Bulletin
of
The Geological Society of America

VOLUME 62 | September 1951 | NUMBER 9

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Subscription $15.00 per year.

Publication Office: Mt. Royal & Guilford Aves., Baltimore 2, Md.

Communications for publication should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117th Street, New York, N. Y.

NOTICE.—Four weeks notice of change of address is necessary. In accordance with the rules established by the Council, claims for non-receipt of the preceding number of the Bulletin must be sent to the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter at the Post-Office at Baltimore, Md., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103, Act of October 3, 1917, authorized on July 8, 1918.

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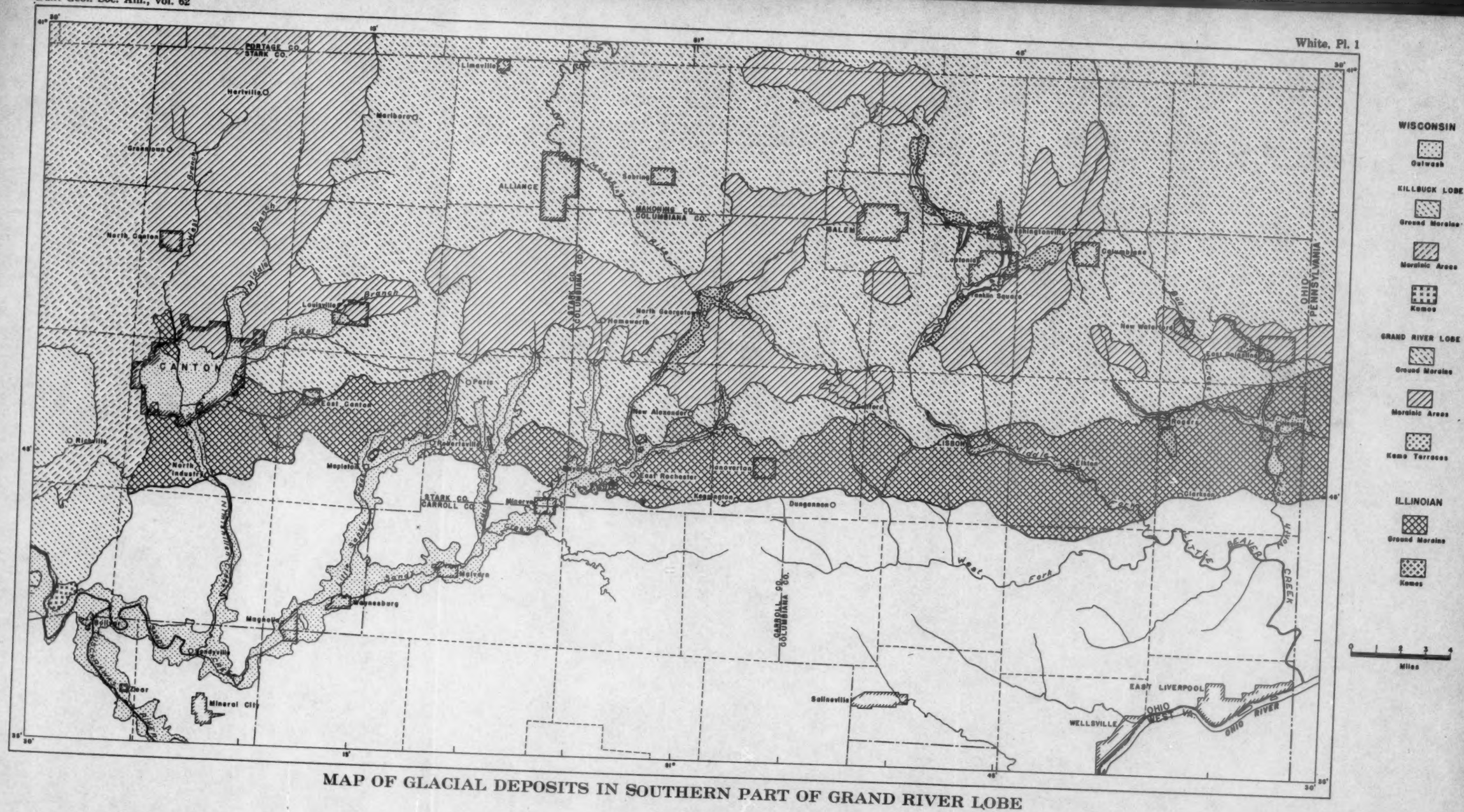

MAP OF GLACIAL DEPOSITS IN SOUTHERN PART OF GRAND RIVER LOBE

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 967-978 2 FIGS., 1 PL. SEPTEMBER 1951

ILLINOIAN AND WISCONSIN DRIFT OF THE SOUTHERN PART OF THE GRAND RIVER LOBE IN EASTERN OHIO

BY GEORGE W. WHITE

ABSTRACT

Glacial drifts of Wisconsin and Illinoian age occur in the southern part of the Grand River lobe in eastern Ohio.

The Illinoian, which occupies a belt only 2 to 5 miles wide, extends eastward from Canton across Stark and Columbiana counties. This is mainly till, now discontinuous and thin. Oxidation reaches a depth of 12 feet or more, and leaching 11 feet or more. Small areas of Illinoian kames and kame terraces remain.

In the Wisconsin till, which covers the area north of the Illinoian, oxidation reaches depths of 8 to 11 feet, and leaching 5 to 7 feet. A partially discontinuous end moraine lies 1 to 2 miles inside the southern limit of this drift, and a very strong end moraine lies along the west side of the lobe.

Within the Wisconsin area relatively minor kame terraces occupy some of the valleys. The Wisconsin ice disappeared mainly by northward retreat of an ice edge, but stagnated to a limited extent in small areas.

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INTRODUCTION

This paper reports (1) the position of the outer margin of the drift across the Grand River lobe in eastern Ohio, (2) the margin of the Wisconsin "terminal moraine" supposed to lie a very few miles north of the outer drift margin, (3) the age of the "extramorainal" drift or "fringe", and (4) the mode of retreat of the Wisconsin ice of the Grand River lobe.

LOCATION

The area studied, in central-eastern Ohio (Fig. 1), is covered by topographic maps of the Canton, Alliance, Lisbon, Columbiana, Dover, Carrollton, Salineville, and Wellsville quadrangles. All of the first four and the northern halves of the last four were used in the construction of the base map for Plate 1.

The area is within the Allegheny plateau

FIGURE 1.—INDEX MAP OF OHIO

Showing Wisconsin glacial lobes, Wisconsin drift border (solid line), Illinoian drift border (broken line) and area shown in Plate 1.

province. The Ohio River crosses the extreme southeastern corner, but most of the area is drained by streams tributary to the Tuscarawas River, by the Mahoning River, or by the tributaries of Little Beaver Creek (Pl. 1). The tops of hills range between 1200 and 1300 feet in elevation. The relief is of the order of 200 feet, but is as much as 300 feet in a few places and less than 100 feet north of Alliance.

ACKNOWLEDGMENT

The field work was principally financed by a grant from the Penrose Bequest of The Geological Society of America, for which appreciation is here expressed.

EARLIER WORK

Early observations on drift in Ohio were made by Volney (1804, p. 83, Fig. 3), Drake (1815, p. 74), and Mather and others (1838). General observations in northeastern Ohio were made by Whittlesey (1866), Newberry (1874; 1878a; 1878b; 1878c), Chamberlin (1883, p. 313), Lewis (1884, p. 200), and Wright (1884a; 1884b; 1884c; 1890). Wright emphasized the "fringe" of attenuated drift in Columbiana

and Stark counties. Leverett (1902, p. 351) interpreted the "extramorainic drift", the "fringe" of Wright, as early Wisconsin and again referred this drift in western Pennsylvania (1934, p. 108) "to the Wisconsin stage." White (1942; Flint and others, 1945) interpreted the outer marginal drift as Illinoian; the present paper sets forth the supporting data. Stout and others (1943, p. 23) referred the "fringe" to a Kansan or pre-Kansan age. MacClintock and Apfel (1944) interpreted the border drift adjacent to the Ohio line in eastern Pennsylvania as Illinoian.

Illinoian Drift

Introduction

Drift interpreted as Illinoian extends from east to west across the area in a belt 2 to 5 miles wide. It lies south of the Wisconsin drift and consists of more or less discontinuous deposits of deeply weathered till, except for three small tracts of kames. It is the marginal drift of an Illinoian Grand River lobe, which came from the same direction and under the same topographic controls as the later Wisconsin ice.

Illinoian Boundary

The boundary of the Illinoian drift extends generally west from the Pennsylvania-Ohio State line across Columbiana and eastern Stark counties to a point 3 miles south of Canton, where the Illinoian is overlapped by the Wisconsin drift of the Killbuck lobe (Pl. 1). This is approximately the southern margin of the "fringe" of Lewis (1884, p. 200) and of Wright (1884a, Pls. 8, 9) from the State line to Robertsville in eastern Stark County, but differs from it west of that village. Across Columbiana County the drift margin as drawn on Plate 1 agrees with that of Stout and Lamborn (1924, map 3).

The boundary is marked by a moraine only near Bayard in western Columbiana County and must be drawn through the locations of the southernmost erratics. At places these are widely spaced and so sparse that the boundary has been shown by a dashed line.

Illinoian Till

The Illinoian till is, in general, so thin that most of even the thicker outcrops consist of only a few feet of strongly altered material between the soil and the bedrock. The rare outcrops of unaltered till exhibit the characteristic five zones or horizons of weathering (Leighton and MacClintock, 1930, p. 31, Fig. 1) similar to those of the eastern side of the Scioto lobe in eastern central Ohio (White, 1939).

Only rarely is the Illinoian drift of the Grand River lobe thick enough to preserve unaltered till. Unaltered till of Horizon 5 is exposed in a strip mine at the southeast corner of the corporation limit of Rogers, in Middleton Township, Columbiana County, as a bluish-gray, moderately calcareous, tough, silty, pebbly to cobbly boulder clay (Fig. 2, A). The top of Horizon 5 is 12 feet below the surface.

Horizon 4 is calcareous till similar to Horizon 5, except for oxidation to a brown color. At Rogers, Horizon 4 is only 8 inches thick (Fig. 2, A), and the depth of oxidation compares well with that in the Illinoian till of the eastern part of the Scioto lobe (White, 1939, p. 165, Fig. 3).

Horizon 3 is quite similar to Horizon 4, except that the carbonates have been leached. The depth of leaching at Rogers is 11 feet, 4 inches. In SW Section 22, Center Township, Columbiana County, the depth of leaching is more than 11 feet, but the base of Horizon 3 is not exposed (Fig. 2, B).

Horizon 2 is the zone of decomposed till underlying the true soil. It is not only oxidized and leached but also thoroughly weathered, and contains fewer pebbles and cobbles than the lower horizons. The upper part of the horizon is mottled buff to gray to brown. The lower part is more uniformly brown and has dark stains along the joints. Horizon 2 has not been weathered to gumbotil, because of the excellent drainage of the region, although in a similar region in New York Illinoian gumbotil has been reported (MacClintock and Apfel, 1944, p. 1147).

Horizon 1 is the soil, divided into A and B horizons of pedologists.

Ground Moraine

The region, underlain by almost horizontal Pennsylvanian strata, was maturely dissected in pre-Illinoian time. It was thinly covered by Illinoian till which at most places only slightly

modified the pre-existing topography. Since Illinoian time the till has been deeply weathered and largely removed. Constructional topography is rarely preserved. The difference in aspect between the area glaciated by Illinoian ice and the unglaciated region to the south is very slight. Far greater difference is exhibited between Illinoian and Wisconsin topography than between Illinoian and unglaciated areas.

Although most small areas of now somewhat subdued constructional topography are related to tracts of kames, a few such areas in which the material is till are in Columbiana County in S Section 9, Middleton Township between Rogers and Negley; small tracts in Center Township, west of Lisbon; near the glacial boundary 1¼ miles east of Dungannon; in Stark County, along the walls of the valley of Little Sandy Creek in Osnaburg Township from 2 miles north of Mapleton northeastward to the township line; in Section 15 and 11 of the same township, northwest of Mapleton; and just southwest of East Canton.

An interesting "island" of Illinoian drift is preserved in the re-entrant angle between the Killbuck and Grand River lobes of Wisconsin drift in the northwestern part of Canton and for a mile north of the city, west of West Branch. Fulton Street crosses the center of the area. The Illinoian drift of the tract varies from thoroughly weathered till rarely more than 2 or 3 feet thick to only scattered pebbles overlying the bedrock which outcrops in many localities along the streets and below the Canton stadium.

Kames

Three areas of kames exist within the Illinoian area. Just southeast of Negley, 1¼ miles west of the State line, gentle kames 10 to 30 feet high lie upon a Teays (Parker Strath) rock terrace (Stout and Lamborn, 1924, p. 41). The material is poorly sorted gravel in which considerable till is intermixed.

From the south margin of Rogers, 5 miles east of Negley, gravelly drift extends southward for about a mile in a valley tributary to Bull Creek. In a coal strip pit a quarter of a mile south of the village thoroughly leached, medium to cobbly gravel 4 to more than 10 feet thick overlies the bedrock. The surface is undulating rather than hummocky and has been much modified by erosion.

The largest area of kames lies along the Illinoian boundary on the south valley wall of Sandy Creek, south of Bayard, just east of the Stark County line, where three irregular tracts each having an area of about a quarter of a square mile occupy re-entrants of the valley wall. Gravel pits expose as much as 40 feet of medium gravel, containing a few cobbles and rare boulders, in beds which dip steeply in various directions. The prevalence of till masses included in the gravel is reported to have caused abandonment of some pits formerly operated. Much of the gravel is leached 8 to 10 feet and oxidized 12 feet or more, but greater depths are not uncommon.

The surface is definitely hummocky; the knolls out from the valley wall rise 40 to 50 feet. Those on the valley wall appear as swells upon the larger bedrock hills. The knolls are smooth and rounded and do not have the sharp, irregular aspect characteristic of Wisconsin kames.

The widely scattered kame areas in the Illinoian drift area of the Grand River lobe are comparable to those of the Illinoian drift area of the east side of the Scioto lobe (White, 1939, p. 169–173).

Valley Trains

The only valley-train deposit referable to Illinoian meltwater is a remnant in the valley of Sandy Creek just east of Magnolia, shown on Plate 1 east of the village and south of the creek as an area without pattern surrounded by later Wisconsin valley train. It has been described by Leverett (1902, p. 285) and by Ireland (1940, p. 1350). An Illinoian valley train was probably once continuous from the drift boundary in Sandy Creek to the Tuscarawas River and thence down that valley. Ireland discovered (p. 1345) an Illinoian valley-train remnant in the Tuscarawas valley about 30 miles south-southwest of Magnolia. Still farther south, along the Muskingum valley, are high terrace remnants which are probably Illinoian, although no detailed studies of them have been reported.

Age Assignment as Illinoian

Proof that the area here mapped as Illinoian is indeed pre-Wisconsin is:

(1) Fresh Wisconsin valley trains rise at or within the Wisconsin drift and pass through the wide valleys of the Illinoian region to the unglaciated area. These trains received no contributions from the Illinoian area. The major erosion of the Illinoian area was accomplished before the fresh valley trains were deposited in the already wide valleys.

(2) The Illinoian till is distinctively more weathered than the Wisconsin till.

(3) The erosion of the Illinoian area has progressed much and distinctively farther than in the Wisconsin area. The erosion has gone so far as to remove much of the Illinoian till which is much more discontinuous and patchy than is the Wisconsin.

This pre-Wisconsin drift is interpreted as Illinoian because the topography and depth of weathering of the till are similar to those of a belt of pre-Wisconsin drift on the east side of the Scioto lobe (White, 1939) whose Illinoian age has been long accepted on the basis of continuous tracing into the wider Illinoian area of western Ohio, Indiana, and Illinois. In both areas erosion has removed much till, drainage is integrated, and the general aspect more nearly resembles that of the unglaciated area to the south and east than it does that of the more recently glaciated Wisconsin drift area. In both Illinoian drift regions the depth of oxidation (base of Horizon 4) is between 12 and 17 feet, the depth of leaching (base of Horizon 3) is between 8 and 12 feet, and the thickness of Horizons 2 and 1 (thoroughly decomposed till and soil) is between 4½ and 6 feet.

Wisconsin Drift of Grand River Lobe

Boundary

The boundary of the Wisconsin drift enters Ohio from Pennsylvania 1 mile south of East Palestine and passes in a general westerly direction to Canton (Pl. 1). At Canton the margin of the Wisconsin Killbuck lobe forms a re-entrant angle with that of the Grand River lobe; the boundary of the Wisconsin drift thence passes southward and southwestward to the Tuscarawas River. The marginal Wisconsin drift of the Grand River lobe seldom exhibits good morainal topography at the actual boundary, although morainal expression 1 to 2 miles north of the boundary is usual.

Till

The Wisconsin till of the Grand River lobe, of which several sections are shown in Figure 1, may be divided into five horizons as is the Illinoian drift. The depth of development is less, and Horizon 2 is much thinner. Horizons 5 and 4 are similar to those of the Illinoian drift.

The unaltered till of Horizon 5 is a blue-gray, strongly calcareous, tough, silty to sandy, pebbly to cobbly boulder clay, in which boulders and blocks up to 6 feet in size may be present.

Horizon 4 is calcareous till oxidized to a brown color, but otherwise similar to that of Horizon 5. The base ranges from 8 to 11 feet below the surface but generally is about 9 feet.

The till of Horizon 3 is noncalcareous, but otherwise closely resembles the till of Horizon 4. The base of the horizon is from 5 to 7 feet, but is generally 6 feet or less below the surface.

The thoroughly weathered till of Horizon 2 is thinner than that horizon of the Illinoian drift. Its base is from 3 to 4½ feet, but usually is about 3½ feet below the surface.

Horizon 1, the A and B soil zones, ranges in thickness from 1½ to 2½ feet and generally is less than 2 feet. The soils are sandy loams, loams, and silt loams of several series, depending on the character of parent material and conditions of drainage under which they developed. The morainic tracts are more characteristically loam, sandy loam, and even gravelly loam in certain kame areas. The soils of the ground moraine are usually silt loam.

Between and north of Alliance and Sebring is the southern tip of an area, extending northward to Lake Erie of very clayey till containing few pebbles and cobbles. The clay till is oxidized 10 feet or more but is leached only 3 to 4 feet, apparently because of tightness and poor internal and surface drainage. The soils developed in this clayey till are fine silt loams,

silty clay loams, and clay loams (Mooney and others, 1915, map; Beck and Gossard, 1919, map).

tent probably contributes to variation in depth of leaching, but its importance cannot be assessed from data from this area.

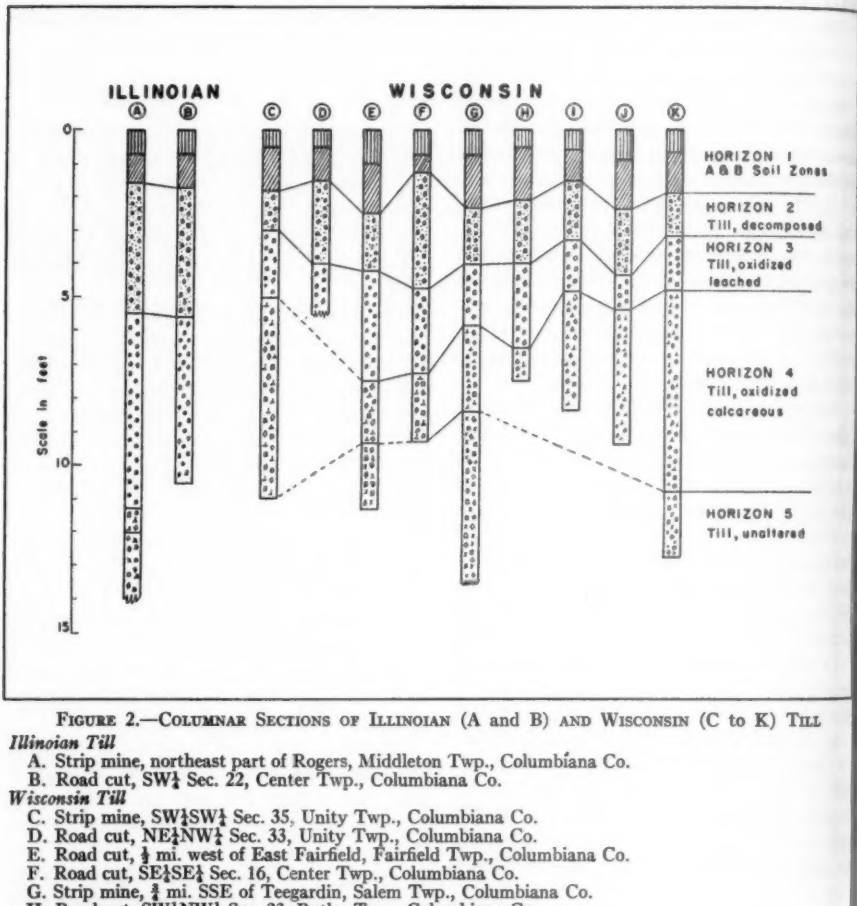

FIGURE 2.—COLUMNAR SECTIONS OF ILLINOIAN (A and B) AND WISCONSIN (C to K) TILL

Illinoian Till

A. Strip mine, northeast part of Rogers, Middleton Twp., Columbiana Co.
B. Road cut, SW¼ Sec. 22, Center Twp., Columbiana Co.

Wisconsin Till

C. Strip mine, SW¼SW¼ Sec. 35, Unity Twp., Columbiana Co.
D. Road cut, NE¼NW¼ Sec. 33, Unity Twp., Columbiana Co.
E. Road cut, ½ mi. west of East Fairfield, Fairfield Twp., Columbiana Co.
F. Road cut, SE¼SE¼ Sec. 16, Center Twp., Columbiana Co.
G. Strip mine, ¾ mi. SSE of Teegardin, Salem Twp., Columbiana Co.
H. Road cut, SW¼NW¼ Sec. 23, Butler Twp., Columbiana Co.
I. Road cut, ½ mi. east of New Alexander, West Twp., Columbiana Co.
J. Road cut, 1 mi. west of New Garden, Hanover Twp., Columbiana Co.
K. Road cut. NE¼SE¼ Sec. 23, Washington Twp., Stark Co.

The considerable variation in depth of leaching in Wisconsin till of the same or only slighly different age is due to a number of factors. Variation in surface drainage has some effect, but permeability of the till is more important. Coarser and more pebbly till is more deeply leached than the finer, more compact till near Alliance. Variation in original carbonate con-

The margin of the clay till, which is not marked by a moraine, is the limit of a late readvance of the Wisconsin ice which had re- treated several tens of miles northward, possibly even into the Erie basin, before readvanc ing with a burden of debris derived from the fine-grained Devonian shales and from lacus- trine deposits of a very early stage of glacia

Lake Erie. This readvance represents a substage of the Wisconsin of northeastern Ohio and is the subject of investigations now in progress (White, 1950).

Ground Moraine

More than half the surface of the Wisconsin drift is classified as ground moraine. It has characteristically a rolling to gently undulating surface. Swells having a relief of more than 10 feet in a quarter of a mile are uncommon except as moraine areas are approached. Irregular tracts of less than 1 to 2 or 3 square miles in Mahoning and northeastern Columbiana counties have a somewhat more undulatory surface than most of the gound moraine and are intermediate in surface character between the ground moraine and the more prominently constructional aspect of the morainic areas. It is not attempted here to map them separately, and they are included with the ground moraine. In the more hilly southern part of the Wisconsin area the drift is plastered on the bedrock hills, producing characteristic undulation of the slopes which thereby differ from the smoother and more regular erosional slopes of the Illinoian and of the unglaciated area to the south.

The ground-moraine surface in the clay till area north of Sebring and Alliance in some places has a relief of less than 20 feet in a mile.

The ground moraine has a variable thickness. At very few places bedrock crops out. Road cuts, strip mines, and well records show that in the ground-moraine areas the drift is usually not over 40 feet thick and very often is between 10 and 30 feet.

Morainic Areas

Included under the term "morainic areas" (Pl. 1) are irregular drift deposits which are probably not genetically homogeneous. The large linear area on the west side of the lobe north of Canton is a massive terminal end moraine. The areas near the south margin are likewise of end-moraine origin. However, parts of these latter areas and other separate areas, especially southeast of Salem, appear not to be true end moraines but masses of hummocky-surfaced thicker drift related to the irregular bedrock surface, especially of valleys and valley walls (White, 1949, p. 34). The designation "morainic areas" is used as a more general term than "end moraine" or "terminal moraine" because the areas are not exclusively such.

One of the two largest areas of well-developed hummocky topography is that along the southern part of the lobe in western Columbiana and eastern Stark counties. Some knolls are as much as 30 feet high and have an area of 20 to 40 acres, but most are not so strong.

In the area from Canton north to Hartville, and extending with even stronger development north of the area of Plate 1 across Portage County to Geauga County, the morainic topography is the most rugged of the Grand River lobe and, indeed, in the whole state. Sharp knolls, many of them kames, are as much as 60 feet high, and farther north in Portage County some knolls are 200 feet high. Near Hartville kettle holes have flat peat and muck-covered floors more than a square mile in area which support an extensive vegetable-raising industry.

The drift of the morainic areas ranges from 20 to 30 feet to well over 100 feet in thickness (Schaefer, White, Van Tuyl, 1946, Pl. 2). Near Hartville a buried valley is filled with 400 feet of drift, but probably not all of this is Grand River Wisconsin drift.

The till of the morainic areas is somewhat coarser than that of the ground moraine. Some of the till, especially that north of Canton, is gravelly, and gravel masses occur in it. The soils are of the Wooster series, mainly Wooster loam and Wooster gravelly loam, but in some of the less well developed areas of moraine, composed of less coarse till, the soils are Volusia (Ravenna) and Canfield loam (Mooney and others, 1915, map).

Kame Terraces

East Palestine area.—Near the headwaters of Leslie Run, a kame terrace an eighth of a mile wide heads in a tributary to Bull Creek, midway between East Palestine and New Waterford in northeastern Columbiana County, and continues downstream on the west and douth side of the valley for 3 miles to East Palestine. The terrace declines from an elevation of 1080 feet at its head to about 1040 feet in East

Palestine. The surface is hummocky with knolls rising as high as 20 feet above the general level and a few shallow kettle holes. The material is gravel with some till included, but details are unknown because of the absence of gravel pits or other good outcrops.

Washingtonville—Franklin Square area.—A complex series of ice-marginal deposits occupies a part of the present valley of Middle Fork of Little Beaver Creek and a part of its earlier valley. The terraces vary in height and were formed at different times. The terrace south of Franklin Square was formed at one time, the middle segment at Washingtonville and Leetonia appears to have been deposited later, followed by still later more northerly deposits. Buried ice blocks must have remained in the axial part of the valley at least as far south as Franklin Square until deposition of outwash ended in the valley; otherwise, a central valley train would be more prominent. It is believed that after the deposition of the terraces the ice masses against which they were deposited shrank, but did not disappear. When outwash no longer came from the north the thinner buried blocks finally melted, and the present valley flat was formed by post-glacial stream action.

From the county line, 1 mile northeast of Salem, a very irregular terrace, made up of massed kames 10 to 40 feet high, lies on the southwest side of the valley. The material is a mixture of gravel, sand, silt in dipping beds, and till masses, bearing evidence of deposition in pools marginal to ice, into which some till slumped and was covered before it could be completely sorted. The contorted bedding indicates collapse as a result of melting of buried and adjacent ice.

Just north of the county line, about 2 miles west-northwest of Washingtonville, a prominent terrace rises sharply 75 feet above the valley bottom. The surface of the terrace ranges in elevation from 1100 to 1120 feet. It is undulatory but has no sharp knolls. A few shallow kettle holes have areas of from 1 to 3 acres. The material is very coarse cobbly gravel, but the character of the bedding is not clearly exposed.

Beginning at the east end of the high terrace a lower terrace follows the north valley side to Washingtonville. Here Middle Fork leaves its wide valley and turns south for 2½ miles in a narrow valley across the upland, rejoining the old valley at Franklin Square. The later narrow valley has no terrace deposits except for low terraces an eighth of a mile wide on either side of the northern half-mile end. The gravel deposits in the abandoned valley continue east through Washingtonville, turn south in the southeast part of the village, and continue to Leetonia. A belt of kames extends southwest from the west margin of Columbiana to join the main line of deposits in southeast Leetonia. The material just east of Leetonia is mainly sand and fine gravel with approximately horizontal bedding making up a level wash plain with an area of 1 square mile.

From Leetonia to Franklin Square only a very narrow and discontinuous kame terrace of coarse cobbly gravel was deposited along the southeast valley wall. From Franklin Square, where Middle Fork re-enters the ancient valley, a strong terrace having a maximum width of half a mile was deposited along the southeast side of the valley for 3 miles southwest. The terrace rises 100 feet above the inner floor of the valley. Its surface declines from 1100 feet near Franklin Square to 1080 feet 3 miles downstream. The surface is markedly hummocky, separate kames are 20 to 40 feet high, and several kettle holes are 10 to 20 feet deep. Two bedrock hills veneered with till rise 60 feet or more above the terrace in the southern part and are large enough to be shown on the map (Pl. 1). At the south end of the terrace the valley of Middle Fork becomes narrow and gorgelike almost to Lisbon; it contains no outwash.

North Georgetown area.—In the vicinity of North Georgetown, in the northwestern part of Columbiana County, very hummocky and irregular kame terraces lie on both sides of the Mahoning River valley and of a tributary entering near Georgetown from the south-southwest. The material is gravel and coarse sand, in which till masses are included. Kames are as high as 40 feet, and some kettle holes are 20 feet deep. The general surface of the terraces in the Mahoning valley declines northwestward from 1220 feet at the southeast to 1160 at North Georgetown. The terraces in the tributary valley are continuous with these in the main valley and decline from 1160 at North Georgetown south-southwest to about 1120

feet at their southern ends, in a direction opposite to the flow of the present stream. The glacial meltwater flowed south to Sandy Creek in a now-abandoned spectacular narrow gorge at the head of the present valley.

Valley Trains

General statement.—Only minor valley-train deposits occur in the valleys of the Little Beaver Creek drainage, but they are much more extensive in the valleys of the Sandy Creek and Nimishillen Creek systems (Pl. 1). The outwash terraces along the Ohio River were not studied.

Little Beaver Creek basin.—In the valleys of Bull Creek and of North Fork near Negley, at the eastern margin of Ohio, are terrace remnants of Wisconsin valley train composed of sand and sandy gravel. The terraces rise from 10 to 20 feet above the present flood plain.

Small terrace remnants of sand and fine gravel lie on either side of the valley of Little Bull Creek at Rogers and upstream from the village. The valley train here is not continuous with that farther downstream at Negley because the valley is very narrow and rock-walled, and no outwash deposits were made in this narrow segment.

A valley train, of which terrace remnants now rise 20 to 40 feet above the present stream, was deposited in the valley of Middle Fork from 2 miles northwest of Lisbon downstream to Elkton. The most extensive remnants are at Lisbon. The material is sand, sandy gravel, and gravel, with occasional larger cobbles.

Elsewhere in the valleys of Little Beaver Creek and its tributaries beyond the Wisconsin border outwash is absent or only scantily represented by a few cobbles and pebbles.

Sandy Creek basin.—The extensive valley trains of Sandy Creek and its tributaries are in distinct contrast to the meager ones of the Little Beaver basin and, indeed, have influenced the very name of the stream. The trains rise at or within the Wisconsin border and pass through the narrow belt of Illinoian drift to the unglaciated area (Pl. 1). The separate trains in the valleys of Upper Sandy Creek, Hugle Run, Little Sandy Creek, and Nimishillen Creek all contribute to the strong valley train of the main Sandy Creek Valley.

The deposits are well-washed and well-bedded coarse sand and gravel. Cobbles up to 4 inches are not uncommon. The soils of the trains are distinctive Chenango gravelly loam. The extent of the deposits and the subsequent erosion of them by present streams within Stark County are well shown on the soil map of the county (Mooney and others, 1915).

Nimishillen valley.—Almost all the area upon which Canton is built is an outwash plain to which deposits were contributed by meltwater which flowed down East, Middle, and West forks to unite in the Canton area. A direct contribution was made from Killbuck ice in the east part of Canton. The outwash passes down Nimishillen Creek and joins the valley train in Sandy Creek valley.

The material varies from boulder gravel at the ice edge in northern Canton to cobbly, medium, and fine gravel away from the ice margin. The outwash from the Killbuck lobe is cobbly near the margin but not bouldery. On the whole the deposits in the Nimishillen system are somewhat finer than in the Sandy system, and the soil is Chenango silt loam (or sandy loam?) rather than gravel loam (Mooney and others, 1915).

Drift of Killbuck Lobe

The Wisconsin drift lobe between the Grand River lobe on the east and the Scioto lobe on the west was originally called the "Shoulder of the Scioto lobe" (Leverett, 1902, p. 403) but is now known as the Killbuck lobe (White, 1934a). The border of the Wisconsin drift of the Killbuck lobe trends south from the junction with the Grand River lobe in the valley of West Branch 1 mile north of the Canton city limit.

The moraine area of the Killbuck lobe is one of striking topographic appearance. As it joins the most western part of Grand River moraine, also of strong development, and as the drifts are similar in appearance, in weathering, in surface expression, and in lithology, it is impossible at present to draw a boundary between the two lobes through the combined moraine tract, and no definite boundary is shown on Plate 1. The more gravelly morainic areas provide large supplies of ground water (Schaefer, White, Van Tuyl, 1946; Kazmann, 1949).

The Killbuck Wisconsin drift is approximately of the same age as the Grand River Wisconsin drift, but evidence now appearing to the north points toward a very slightly later time of maximum advance of the Killbuck lobe.

Age of Wisconsin Drift

The Wisconsin drift must be correlative with either Tazewell or Cary of the Mississippi valley succession, but the age cannot be established solely from this area. Work now in progress to the north and northwest, where correlation can be made with definite end moraines entering the plateau from the till plains, indicates that the margin of the clay till of northeastern Ohio, which enters the area of Plate 1 for only a short distance near Alliance, is pre-Defiance but apparently only slightly so and is therefore middle to late Cary. There is indication 20 miles north of the northwest part of this area that the time between retreat of the Grand River ice and the readvance to deposit the clay till was not long. As the clay till is middle to late Cary, the coarser, earlier Wisconsin drift of this region is probably early Cary. It is believed not to be Tazewell, as Wisconsin drift near Akron is definitely older than the Grand River drift of the area here discussed and is probably Tazewell (White, 1950).

Summary of Glacial History

Pre-Illinoian

No evidence of pre-Illinoian glaciation exists within the area described. Indirect evidence of pre-Illinoian glaciation elsewhere in northern Ohio is afforded by the sequence and pattern of drainage changes (White, 1934b), but these have not been traced into the present area.

Illinoian

The demonstrably older drift south of the fresher Wisconsin drift was deposited by a Grand River lobe of Illinoian ice which invaded northeastern Ohio and northwestern Pennsylvania. In its advance the ice deposited over the bedrock irregularities a till sheet which near its margin was certainly thin, and may have been discontinuous. No strong terminal end moraine was formed, but in a few widely separated favorable places marginal kames were left. Retreat of the marginal ice was at least in part by stagnation, as minor kame terraces are preserved near Negley and Rogers.

Meltwater flowed to the Ohio River, and valley trains, of which only widely separated small remnants now remain, were deposited beyond the glaciated area.

Wisconsin

The Wisconsin ice advanced from a generally northerly direction in a Grand River lobe to the margin shown on Plate 1. A terminal moraine was not made precisely at the farthest limit of ice advance, but an end moraine several miles wide, not entirely continuous from east to west, was formed near the limit of advance. An east-west moraine over 6 miles long north of Salem indicates at least local readvance and oscillation along a line 10–12 miles north of the line of greatest advance.

After the farthest advance, the retreat of the ice to the northern limit of the area was in part irregular in that the ice melted down as well as back. Ice tongues and masses were left in some of the valleys. These may in part have been projections of the main ice sheet, but some of them were entirely separated. Over, around, and at the sides of these elongate masses kame terraces were deposited. The absence of kame terraces in some valleys where they might be expected shows that not all valleys retained dead ice masses. The ice dissipation appears to have been not entirely, or even in large part, by stagnation but by northward retreat of generally active ice, but which had in some places areas of weakly active or completely dead ice.

From the time of maximum ice advance and thereafter meltwater escaped southward, laying down valley trains which extend far beyond the area.

In the Killbuck lobe the ice advanced from a generally northwesterly direction to a north-south line through the west part of Canton. The advance of the Killbuck lobe was at about the same time as the Grand River lobe but may have been slightly later. The margin in contact with the Grand River lobe is not definite

From the Killbuck lobe was deposited a strong marginal moraine, in part a kame moraine. So little of the Killbuck lobe is included in the area that discussion of the mode of ice retreat of that lobe is not appropriate here.

Meltwater from the Killbuck lobe flowed down West Branch and east from the ice margin in western Canton to that stream and thence down Nimishillen Creek to Sandy Creek. Meltwater from the more southerly Killbuck lobe flowed directly to the Tuscarawas River.

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Manuscript Received by the Secretary of the Society, November 15, 1950

Project Grant 363-41

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 979-1016, 16 FIGS., 6 PLS. SEPTEMBER 1951

PETROLOGY OF THE BALTIMORE GABBRO, MARYLAND

BY NORMAN HERZ

ABSTRACT

The Baltimore gabbro, made classic by G. H. Williams, is found to conform to modern petrologic theories. Two plagioclases, An_{78-88} and An_{64-70} distinguish themselves. The mafic minerals, especially orthopyroxene, show evidences of strong fractionation. Alleged primary brown poikilitic hornblende of Williams and others proves to be a product of uralitization, and primary hornblende is distinguished solely by texture. On this basis, a new rock type, bojite, is defined. The structure of the gabbro complex conforms with an origin in magmatic action, and many structures previously alleged to be metamorphic prove to have an igneous origin. The general structure is that of a funnel with the southeast lip dipping less steeply than the northwest. On the basis of iron enrichment, three distinct periods of crystallization are described: an ultramafic, a gabbroic, and a bojitic. Respective conditions prevalent in the magma chamber are discussed. Differentiation trends of Maryland rocks show absolute enrichment in iron throughout most of the gabbro complex crystallization; alkali enrichment begins with the bojites and is emphasized later in the quartz-bearing rocks. Alteration processes at work in the complex seem to bear out the laboratory system MgO–SiO_2–H_2O of Bowen and Tuttle and analogies are drawn between the two. The sequence of intrusion is believed to be: ultramafic border rocks, gabbros with included ultramafic bands, and quartz-bearing rocks.

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Introduction

The Problem

The Baltimore gabbro complex has intrigued geologists since the days of George Huntington Williams. It comprises an area of about 50 square miles in the Piedmont of Baltimore City and County and Howard County of Maryland. No worker has seriously re-examined the petrology of Williams' 1886 study. This led to the conclusion that the area is an exception to the usual sequence of intrusion, ultramagnesian to quartz-bearing, because Williams considered that the ultramagnesians intruded the gabbros. The construction of the Baltimore-Patapsco aqueduct exposed a complete cross-section through the gabbro complex (Herz, 1950) and its relationships to the country rock, making this present study possible.

Acknowledgments

The writer expresses his appreciation to Dr. Ernst Cloos for suggesting the problem and for his constant encouragement. Dr. J. L. Anderson, Dr. R. W. Chapman and Dr. J. T. Singewald, Jr., offered invaluable criticism and assistance. Dr. J. W. Peoples and Dr. C. B. Sclar read and criticized the manuscript. Mr. B. L. Werner, Resident Engineer and Mr. Cameron Cromwell, Senior Associate Engineer, for the city of Baltimore and their staffs made engineering plans and equipment available, and advised on underground procedures. The authors' wife, Rhoda Herz, drafted many of the illustrations. The Veterans' Administration financed the study.

Geologic Setting

Southeast of the gabbro, sediments of the Coastal Plain cover the Piedmont. The Baltimore gneiss domes dominate the vicinity, were the earliest determinants of structural features, and have delineated the gabbro area.

This gneiss is universally accepted as Precambrian. Uncomformable on the gneiss is the Glenarm series, of metamorphosed sediments, probably early Paleozoic, and in descending order:

Peach Bottom Slate
Cardiff conglomerate
Peters Creek schist
Wissahickon schist
Cockeysville marble
Setters quartzite

The gabbro intrudes the Wissahickon and Peters Creek (considered as one on Figure 1) and is cut by quartzose intrusives, including the Relay quartz diorite, the Port Deposit granodiorite gneiss, and the Ellicot City Granite. The Relay occurs persistently along the southeast margin of the gabbro.

Evolution of Ideas

Williams (1886 and later) distinguished two gabbros intimately associated, a pyroxene gab-

FIGURE 1.—SETTING OF THE GABBRO COMPLEX
Modified from Knopf and Jonas (1929).

bro and a "gabbro-diorite"; a third, peridotite or olivine gabbro, was generally found in small dikes. He assigned a Precambrian age on the basis of occurrence with alleged Precambrian schists and gneisses. The hornblende of the "gabbro-diorite" was always a green "para-

morph" from pyroxene. A brown hornblende in both types was considered primary. The peridotites and pyroxenites were derived from the gabbroic rocks and were younger because of their position in the west and the westerly dip of the crystalline schists (1890, p. 37). Olivine in these rocks crystallized after pyroxene because of its advanced serpentinization.

Leonard (1901) proposed metagabbro for gabbro-diorite. He described a diorite with primary green hornblende which graded into quartz diorite.

Bascom (1902) considered a granite of medium acidity to be the first intrusion because gabbro intruded it. Grimsley (1894) had called this gabbro a segregate.

Knopf (1921) suggested the serpentinized peridotite series to be a "floor phase" derived by a gravity differentiation. She emphasized the southeast dip of the schists to the west of the gabbro and the occurrence of serpentines on the northwest border of the intrusion.

Insley (1928) regarded the ultramagnesian rocks as gradational from the gabbro magma. He attributed uralitization and serpentinization to dynamic and hydrothermal metamorphism, following the work of Williams and others.

Knopf and Jonas (1929) had the ultramagnesics younger than the gabbro and injected them into the country rock, as at Soldiers Delight and elsewhere. They explained ultramagnesics within the gabbro proper as local unwarpings of floor. Pyroxenites could have formed when pressure was relieved on the "crystal raft" of half liquid, half solid olivine mush by re-melting the olivine. The resulting magma might attain the composition of a pyroxenite. Peridotites could not be explained in such a way.

Cloos and Hershey (1935) assigned a Paleozoic age to the gabbro complex. Tracing from known Paleozoic rocks, along the Susquehanna River, they found Paleozoic structures engulfed by the gabbro, thus making the latter at least post-Conestoga (lower Cambrian) and perhaps Appalachian.

Cohen's structural study (1937) will be reviewed on another page.

Mineralogy

General Statement

The most important primary minerals of the Baltimore gabbro are plagioclase, pyroxenes, and olivine. Uralitization and saussuritization processes raised colorless amphiboles, hornblende, epidote, and chlorite to importance; serpentinization of ultramagnesian types produced serpentine, chlorite, colorless amphiboles, and chromite.

Methods of Study

Indices of Refraction.—By immersion methods, using white light, refractive indices were determined, for a time with oils calibrated in steps of .01 index; later in steps of .004. The absolute index of the oils of the .01 series was frequently checked with an Abbe refractometer. Estimated possible errors in the early work are ± .003; the later work ± .001.

For the plagioclases, indices on the (001) or (010) cleavage plates were determined. Gamma index of orthopyroxene was determined from the (110) prism. Clinopyroxene (100) parting plates gave a value close to beta (yield a slightly off-center optic axis figure, about 20°). Olivine beta is parallel to crystallographic c and was obtained from sections showing c. Amphibole (100) yielded a slightly off-centered optic axis figure giving a value close to beta.

Optic Angle.—All optic angle measurements were made on a Leitz four-axis universal stage. Because of the generally large optic angle in the orthopyroxenes, olivines, and amphiboles, measurements by direct rotation from one optic axis to the other were less common than measurements from an optic axis to a bisectrix. Hess (1949) states that measurements by direct rotation, as was usually done for the clinopyroxenes, are better than ± 0.5°; by measurement to a bisectrix, errors of 5° can occur.

In this work, the error in recorded 2V is probably under 5°, since several observations (usually six) were made on each slide, and then averaged. The different observations generally agreed to within 5°; the probable error is ± 2°.

Extinction Angles.—Extinction angles were determined on the universal stage, using the

method of Burri (Haff, 1941). The accuracy of these measurements is less than ± 2°.

Other Techniques.—Many powders to be used in index determinations were separated on the Franz Isodynamic Separator. Minute quartz fractions in altered gabbros were occasionally found this way.

A Berman balance gave specific gravity determinations on all garnet samples to an accuracy varying as the purity of the sample.

An integrating stage gave modal compositions with a probable error of about ± 2%.

Plagioclase

Plagioclase usually constitutes over half the volume percentage of most types except the ultramagnesian. It is generally of irregular shape, rarely lath-like. Grains are sometimes very large when associated with a green amphibole, indicating a later pegmatitic development; otherwise grain size is comparable to the associated pyroxenes. Immersion methods gave the albite: anorthite ratio using the curves of Tsuboi (1925). In restudying the slides of G. H. Williams, the migration curves of Federow (Emmons, 1943), and Turner (1947) were used. Twinning was determined entirely with these curves.

The total variation in composition is $Ab_{14}An_{86}$ to $Ab_{68}An_{32}$. A statistical study of these determinations revealed a definite concentration of certain compositions (Table 1).

The plagioclases represented are bytownite, mean composition $Ab_{30}An_{70}$, 67 per cent, and labradorite $Ab_{34}An_{66}$, 33 per cent. Plotting these compositions against distance from contact (Fig. 2) reveals no systematic relationship. Though the most calcic ($Ab_{14}An_{86}$) was only 600′ from the schist contact, other determinations revealed no tendency for its crystallization near the contact where the ultramagnesics are largely localized.

The more sodic plagioclase tends to be associated with uralitized mafics. In every example of group V range (Table 1), the primary mafics were either all gone, or left as scattered remnants. In this range, saussuritization of the plagioclase indicated that crystallization as a primary magmatic feature did not continue far beyond group IV.

There is no systematic variation between plagioclase and orthopyroxenes (Fig. 3). Group II started crystallizing when the orthopyroxene was more magnesia-rich than did group IV. No plagioclase occurs where orthopyroxene is more magnesic than En_{86}, and there it is $Ab_{17}An_{83}$. The initial plagioclase of group IV $Ab_{32}An_{68}$ appears at En_{79} and once group IV appears the two plagioclases crystallize out together. Direct evidence on the later appearance of group IV is the "intrusions" by this group into group II and contributing to the uralitization of the associated pyroxenes (Pl. 2, fig. 2).

TABLE 1.—PLAGIOCLASE COMPOSITION CONCENTRATIONS

Group No.	Composition Range	Number of Determinations	Percentage of Whole
I	An_{84-86}	4	7
II	An_{78-88}	28	51
III	An_{71-77}	5	9
IV	An_{64-70}	12	22
V	An_{63-68}	6	11
		55	100

Thus, there is no complete mixed crystal series in the plagioclases as occurs in the mafics (see below). Group I leads abruptly into the most important of all, group II, of composition An_{78-88}. Group III is a transition group leading into IV, An_{64-70}. Group V is probably developed from alteration. This picture of a grouping indicates that strong fractionation of the plagioclase must have taken place only at very definite times during magmatic crystallization.

Twinning laws were determined with the hope of finding some sort of systematic variation in the different groups (Table 2). Generally the calcic plagioclases tend to twin according to the so-called less common laws; with increasing soda, the albite law assumes importance. Mannebach twins are universally present.

Well developed opaque inclusions are occasionally seen in plagioclase. Williams (1886) described them, but offered no explanation of

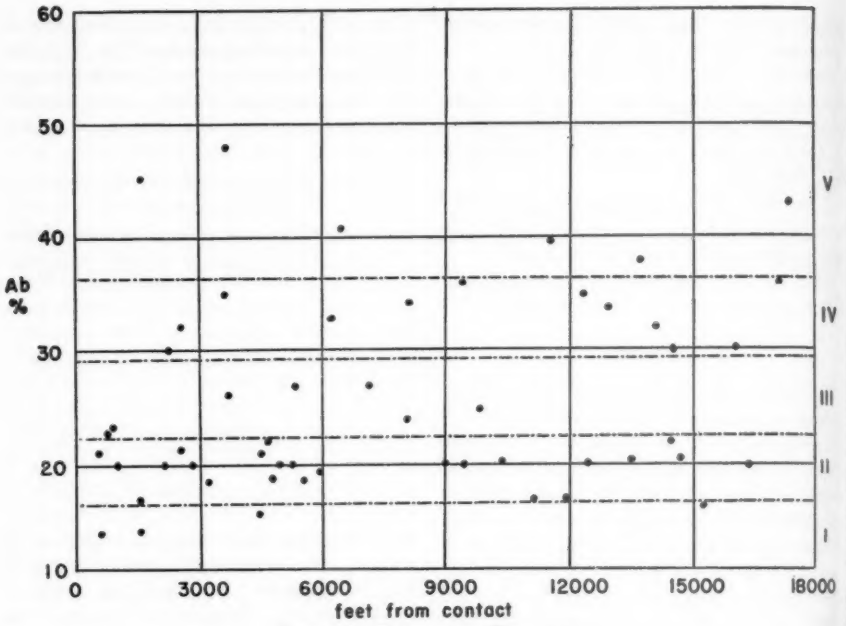

FIGURE 2.—PLAGIOCLASE VARIATION

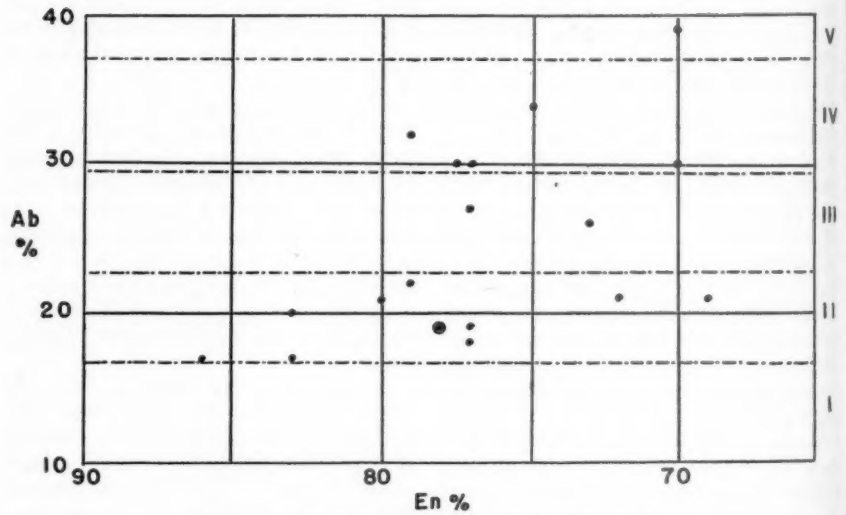

FIGURE 3.—PLAGIOCLASE/ORTHOPYROXENE RELATIONS

their chemical or mineralogical nature:

> The needles occupy the center of the crystal; this is surrounded by a zone where only the dots or globulites are to be seen. The exterior of the crystal is generally free from inclusions of any kind. Frequently the acicular bodies are altogether absent and then the minute dots are arranged in lines which indicate the position of the twinning lammelae (p. 21).

Thin-sections reveal nothing of the nature of these inclusions; the Franz separator distinguished plagioclase with inclusions from clear plagioclase. The least magnetic fraction is invariably more sodic and free of inclusions; the more magnetic more calcic and containing the inclusions. Although the separator cannot be calibrated in terms of absolute units, it does appear that iron must play an important role.

TABLE 2.—PLAGIOCLASE TWINNING LAWS

	Plagioclase group		
	II	IV	V
Normal Twin Laws			
Albite	R	A	A
Mannebach	A	P	P
Parallel Laws			
Carlsbad	R	R	M
Acline	A	M	M
Pericline	P	M	M
Complex Laws			
Albite-Carlsbad	R	M	M
Albite-Ala	P	A	M

A abundant, P present, R rare, M missing.

Evidences of syncrystalline deformation are common. Bent twinning lamellae are universal in occurrence as are incomplete or exsolution lamellae. Granulated areas occasionally are seen with pyroxene, usually altered to amphibole, but with plagioclase rarely changed.

Well developed reaction rims occur especially where magnetite or olivine is also present typically:

W54 — olivine — orthopyroxene, .008–.015 mm. — clinopyroxene, .00–.03 mm.—amphibole-spinel symplectic, .07–.10 mm.—(garnet?, .003 mm.)—plagioclase.

W69 — olivine — clinopyroxene, .02–.03 mm. — amphibole-spinel symplectic, .07–.08 mm.—plagioclase.

C459-1 — magnetite — amphibole, .005–.300 mm. —plagioclase.

Precise optical data on these corona minerals could not be obtained but extinction angles and pleochroism helped identify them. Clinopyroxene was irregular, more or less equidimensional, with no common orientation; extinction averaging 41°. The amphibole symplectic was fibrous and vermicular with orientation varying in each fiber cluster; extinction was about 15°, pleochroism from colorless to green-blue. Orthopyroxene was aligned in ragged prisms with c axes perpendicular the olivine and with lower birefringence than the clinopyroxene (Pl. 4, fig. 2). Where amphibole comprised the entire corona, it was usually more strongly pleochroic, from bluish-green to pale green, fibrous or irregular, and with an extinction of about 20°. Pale amphibole with lower extinction may be actinolite-tremolite; the more pleochroic, hornblende. Occasionally, rimmed plagioclase borders serpentine, but olivine remnants indicate the reaction to pre-date serpentinization

TABLE 3.—OPTICAL DATE ON OLIVINES

Specimen	2V	Y index	Estimated Fo content	Estimated Related En
T108-2	+84° to +86°	1.660	96%	98%
T80-1, T114-1, T127-1	+87° to 90°	1.675 (av)	89%	88–91%
C366-1	−86° to −87°	1.694	78%	86%
T105-1	−81° to −84°	1.711	70%	77%

Plagioclase is often altered by albitization, epidotization, and saussuritization, but is, however, stabler than the mafics. Thus uralite occurs with all the plagioclase groups. Alteration sometimes begins with the development of clinozoisite scattered throughout the crystals, other times along the edges in contact with uralite (Pl. 3, fig. 3). In earliest alteration it is irregular; later it assumes a prismatic shape. In advanced alteration, clinozoisite tends to crystallize between plagioclase and uralite, indicating a mutual reaction for its formation. That the mafic mineral did play a role is indicated by the average clinozoisite determination revealing 7 per cent of the epidote molecule ($HCa_2Fe_3Si_3O_{13}$). With further alteration, plagioclase becomes more albite-rich, forming albite laths; carbonate and colorless amphibole may appear in very advanced stages.

Olivine

Olivines were determined with the curve of Winchell (1933) and checked with that of Deer and Wager (1939); except for the slides of Williams, optic angle and the median index were determined (Table 3).

Olivine is abundant in the ultramagnesian rocks and occurs occasionally in the gabbroic. It is usually in small, rounded crystals, though grain size varies as the composition. Average determinations show:

Forsterite 96%	0.4 mm.
94%	1.0
90%	1.5
89%	1.2-2.0
78%	0.6
70%	0.4

Composition ranges from $Fo_{96}Fa_4$ to $Fo_{70}Fa_{30}$, with the largest grain size about $Fo_{89}Fa_{11}$. The significance of this composition is also brought out by the fact that 70 per cent of all olivine determinations fall in Fo_{88-90}.

As expected, there is a definite relationship between orthopyroxene and olivine crystallization (Fig. 4). As orthopyroxene becomes richer in iron, olivine does the same.

Zoning was not observed in olivine. This may be due to the universal serpentinization of the olivines at least about their margins, destroying any evidence of the phenomenon. Olivines of more than one composition, however, occasionally occur in the same thin section. Thus, T86-1 has one group of 2V +84° to +86° (Fo_{97}); another 2V −89° (Fo_{88}). C457-4 has a 2V of +84° to +88° (Fo_{93}) and −86° (Fo_{89}).

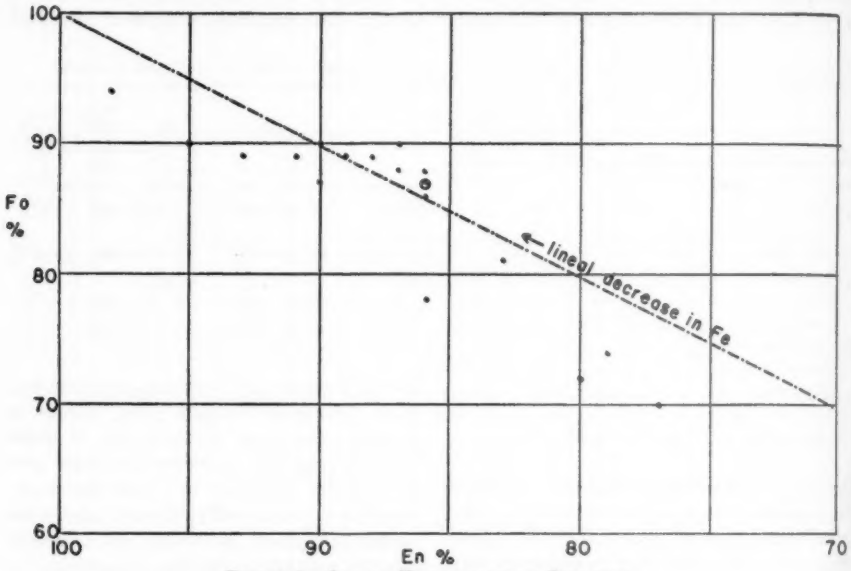

FIGURE 4.—OLIVINE/ORTHOPYROXENE RELATION

It appears that complete reaction between the crystals and melt was sometimes prevented and that later serpentinization destroyed the armor plating of the earlier formed crystals.

Olivine is the mafic mineral most susceptible to hydrothermal alteration. This is in the form of serpentinization, beginning along cleavage cracks and margins of the grains, and always present to some degree. The serpentine is fine fibers or laminae of antigorite, showing a weak birefringence, colorless to pale green. Coincident with this formation is magnetite, also forming in cleavages and on margins. In rocks that have been nearly completely serpentinized, magnetite often delineates the outlines of the former olivine crystals (Pl. 1, fig. 4).

Orthopyroxene

The more magnesian orthopyroxenes were determined with Hess and Phillips' curve (1940); the more iron-rich by Poldervaart's (1947). Optic angle was determined in every case; gamma index as a check (Table 5).

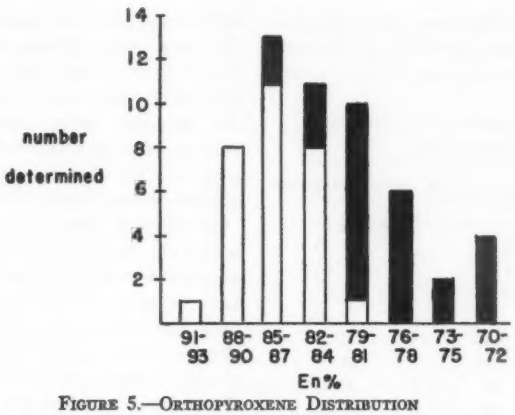

FIGURE 5.—ORTHOPYROXENE DISTRIBUTION

55 determinations. Black = gabbros; white = ultramagnesics.

TABLE 4.—ZONING IN ORTHOPYROXENES

Specimen	2V		Rock type
	Core	Margin	
T14-2....	+52°	+73°	Magnetite-rich gabbro, En_{89} most common
C374-1...	+69°	+81°	Normal gabbro, En_{75} usual orthopyroxene
C395-1...	+56°	+61°	Pyroxenite, over 75% orthopyroxene En_{84}
C459-1...	+56°	+60°	Normal gabbro En_{75} usual orthopyroxene

TABLE 5.—OPTICAL DATA ON ORTHOPYROXENES

Specimen	2V	Gamma index	Estimated En Content	Related	
				Fo	Ab
T8-1....	+60° to +66°	1.676	91		
T86-1...	+85° to +88°	1.677	90	90	
T127-1..	−88°	1.679	88	89	
C401-4..	−66° to −72°	1.692	78		27
C376-1..	−63°	1.701	73		26

Orthopyroxene is almost a constant constituent, except in strongly uralitized rocks. In the ultramafics it tends to be phenocrystic, often enclosing other minerals. Grains up to 10 mm. in diameter are not uncommon where orthopyroxene dominates. The more feldspathic rocks are roughly equigranular with grain size ranging from one half to 2 mm. Pleochroism is generally lacking but in Williams' thicker slides it was easily discerned as X = brownish-red, Y = reddish-yellow, and Z = green.

Optical determinations indicate a range from $En_{99}Of_1$ in a zoned crystal, possibly to $En_{50}Of_{50}$ (Of = orthoferrosilite) and two different orthopyroxenes involved in the crystallization. The early, more magnesium one En_{95-99} (enstatite) exists as zoned crystals only, and is volumetrically insignificant. The dominant orthopyroxene En_{70-93} (largely bronzite) is later and more iron-rich; two determinations on Williams' slides indicated En_{68} and En_{50}, but these could not be checked with index oils.

The distribution of this dominant group (Fig. 5) indicates that it grades as an entity from ultramafic rock varieties, with a magnesian-rich pyroxene, into gabbroic, with an iron-rich one. Thus, unlike the olivines the pyroxenes are often found zoned. In every case an increase in optic angle indicates the core more magnesium-rich than the margins and a systematic enrichment in iron taking place from core to rim (Table 4).

The ultramagnesian rocks' orthopyroxene is concentrated largely within En_{83-90}; only two determinations fall outside this range (Fig. 5). Coincidently, olivine as has been pointed out, appears concentrated within Fo_{86-90} (70 per cent of all determinations).

Orthopyroxenes of more than one composition often occur in a single slide; an ultramag-

nesian zoned, or reaction rimmed crystal together with a more iron-rich type (Table 4). The early zoned crystal is invariably larger than the average grain size of its host rock. In C459-1, the magnesian orthopyroxene averaged 1.2 mm. other minerals, 0.7 mm. It appears then that complete reaction between early formed crystals of orthopyroxene and olivine, and the melt was often prevented. Kuno (1936) and others have also noted zoned orthopyroxene phenocrysts to be more enstatite-rich than unzoned types, so the phenomenon is not unique.

TABLE 6.—CHEMICAL ANALYSES OF ORTHOPYROXENES

	Orthopyroxene from Uralitized Gabbro	Orthopyroxene from Pyroxenite
SiO_2	52.12	54.53
Al_2O_3	1.69	1.93
FeO	20.94	8.92
Fe_2O_3		1.70
CaO	3.20	2.25
MgO	21.36	29.51
Cr_2O_3	nd	0.30
MnO	nd	0.28
H_2O	nd	1.14
	99.51	100.56
Specific gravity	3.35	3.300
Formula	$En_{60.8}Fs_{32.9}Wo_{6.4}$ or $En_{64.9}Fs_{35.1}$	$En_{79.5}Fs_{16.2}Wo_{4.3}$ or $En_{82.1}Fs_{16.9}$
Analyst	W. G. Brown	T. M. Chatard
Reference	Williams 1886	Williams 1890

Inclusions within the orthopyroxene are quite common. Williams, who first described them regarded them as brookite (1886, p. 24):

> They are composed of fine needles arranged parallel to all three crystallographic axes and of little plats of a reddish-brown color lying parallel to the brachypinacoid. They vary considerably in thickness, as may be seen by the different depths of color which they possess. The plates all extinguish simultaneously with the hypersthene, which may be due to their extreme thinness or the fact that they have their axes parallel to those of their host.

These rod-like inclusions most often parallel to the c crystallographic axis, sometimes either a or b (Pl. 4, fig. 1). Smaller plates (schiller structure) are common parallel (010), as stated by Williams; but also parallel (100) and (110). The exact nature of these diminutive inclusions cannot be determined since their properties are masked by their host. Chemical analyses do not rule out their being brookite; average of 9 analyses of rocks with orthopyroxenes revealed 0.08 of TiO_2 though neither orthopyroxene analysis showed titania (Table 6).

In addition, there are irregular, exsolution inclusions, probably of diopside, in the more magnesian orthopyroxenes. These invariably have the same crystallographic orientation to their host with both c axes coincident, and orthopyroxene a equal diopside b; thus, the optic plane of clinopyroxene parallel to orthopyroxene (100) (Fig. 6). Extinction angles of the inclusions vary from 34° to 43°, averaging 40°, thus closer to the diopside end of the diopside-hedenbergite series. This orientation is identical to the diopside exsolution plates described by Hess and Phillips (1940) at lower temperatures and in a more iron-rich environment.

Diopside exsolution plates were rare and occurred in orthopyroxenes of composition about En_{70} (Pl. 1, fig. 1). The plates are colorless averaging about 45° in extinction; 2V determinations are highly erratic.

The rare occurrence of these diopsidic lamellae forms an invaluable geologic thermometer. The final stages of crystallization have been just below the inversion curve ortho-clinopyroxene (Fig. 7) intersection with the magmatic temperature curve, slightly over 1100°C. (Hess 1941, p. 582). Also, since orthopyroxenes never show any evidence of inversion from monoclinic forms, magmatic temperatures were probably never above 1140°C. (Fig. 7; Bowen and Schairer, 1935, p. 200).

Orthopyroxene reaction rims are common, especially where the contact is with magnetite or pyrite. Coronas similar to those observed in the plagioclase-olivine contacts are also seen at orthopyroxene-plagioclase contacts. Since the only products within these coronas, however, are amphiboles (Pl. 3, fig. 1), these may be uralitization phenomenon, akin to hydrothermal alteration and not reactions in the solid state.

Orthopyroxenes are highly susceptible to alteration by both serpentinization and uralitization. Serpentinization usually proceeds along

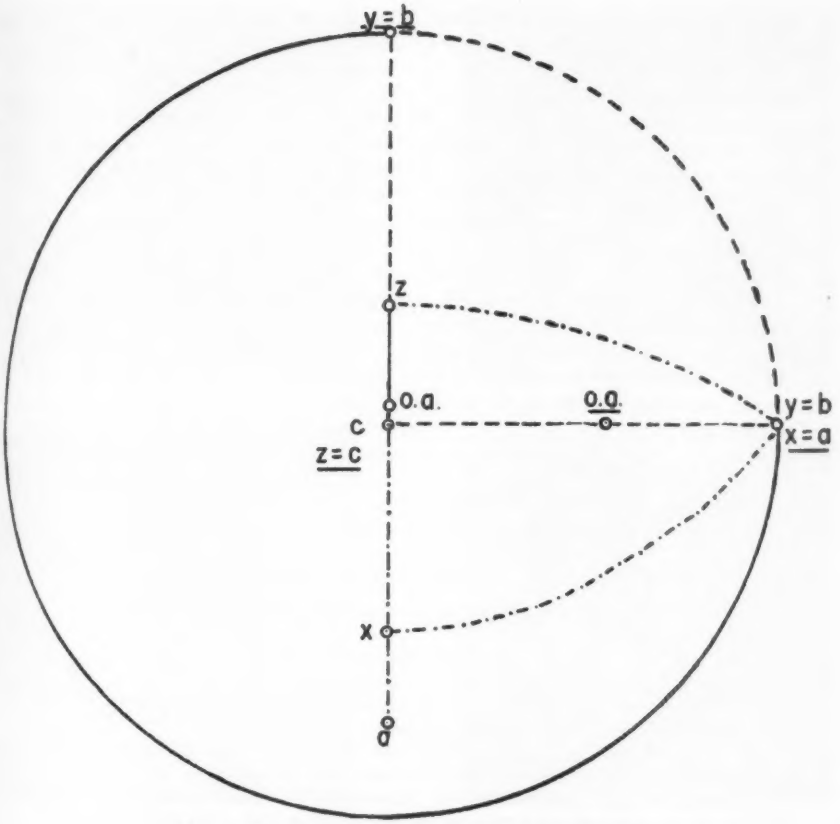

FIGURE 6.—DIOPSIDIC INCLUSIONS IN ORTHOPYROXENE
Orthopyroxene underscored.

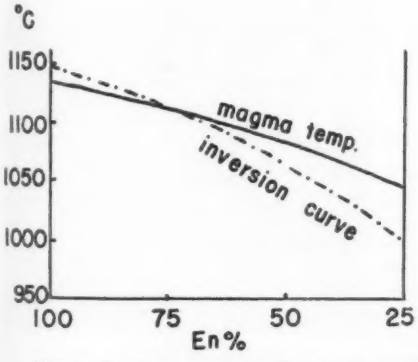

FIGURE 7.—ORTHOPYROXENE INVERSION CURVE
After Hess (1941, p. 583).

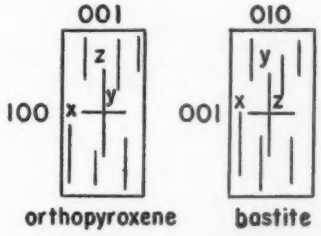

FIGURE 8.—ORTHOPYROXENE/BASTITE OPTICS

cleavages in the form of an antigorite, bastite, and different stages can be seen on different slides (Pl. 1, fig. 3). In this, the (110) cleavage of the enstatite is preserved, but with serpentine Y parallel to its trace, taking the place of ortho-

pyroxene Z. The acute bisectrix, X (equals crystallographic c), of the serpentine is then perpendicular to this cleavage plane (Fig. 8).

Alteration to uralite reveals no symmetrical change, though different slides show that it preceded serpentinization. The first alteration to tremolite and sometimes anthophyllite is also indicated by the tendency of these amphiboles to form the innermost corona in incipient uralitization, arranged in discrete fibers or prisms perpendicular to the edge of the pyroxene. The outermost corona is invariably a pleochroic green hornblende, usually fibrous but occasionally compact. The presence of magnetite appears to heighten the development of the hornblende, increasing both its depth of color and its total area.

Clinopyroxene

The optic angle and Z∧c were determined in every case; beta refractive index was determined when the rock was available. Chemical composition determinations from optical data proved difficult (Table 7): the method decided upon was the intersections of observed beta index with the curve of normal trend of clinopyroxene crystallization (Hess, 1941). Walker and Poldervaart (1949) used the gamma intersection with this curve; Hess (1949, p. 636) suggests the intersection beta with the curve as a check on compositions inferred from 2V and the beta index. The significance of this trend curve is indicated by the two chemical analyses falling exactly on it (Fig. 9). The scattering of points plotted by optical data is also shown in the diagram.

Clinopyroxene is the second most abundant mineral in the gabbros (after plagioclase) and about equal to the orthopyroxenes in the ultramafics, where it tends to be phenocrystic, ranging up to about 8 mm. in diameter. In gabbroic types, which are equigranular, it varies from

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PLATE 1. PHOTOMICROGRAPHS

FIGURE 1.—ORIENTED DIOPSIDIC PLATES IN ORTHOPYROXENE IN GABBRO
T19-1, plane light, ×21.

FIGURE 2.—HYPIDIOMORPHIC BOJITE
C336-1, × nicols, ×21. Note clear, unaltered plagioclase twins.

FIGURE 3.—SERPENTINE BASTITE (sp) DEVELOPING FROM ORTHOPYROXENE (p)
T80-1, plane light, ×21. Note continuous cleavage.

FIGURE 4.—SERPENTINE ANTIGORITE DEVELOPING FROM OLIVINE
T80-1, plane light, ×21. Magnetite appears to outline former crystals.

FIGURE 5.—MICRO-BOJITIC FILLING IN UNDEFORMED GABBRO
C445-1, × nicols, ×21. Note large pyroxene (p) and plagioclase (pl) in the micro-bojite.

FIGURE 6.—BOJITE INTRUSION INTO ULTRAMAFIC
Hollofield, plane light, ×7. Note development of large garnet (g) in ultramafic; bojite comprises rest of slide.

Note: slides are designated as follows—T, from the City section of the aqueduct; C from the County section. The number preceding the dash is the hundred foot station; following the dash, the order collected in that hundred foot station. Thus, C336–1 is the first specimen collected in the 100′ county station from 33,600′–33,700′, measured from the western end as 0′. The city stations are measured from the city line as 0′ (Fig. 1).

PLATE 2. PHOTOMICROGRAPHS

FIGURE 1.—GRADATIONAL CONTACT BETWEEN PYROXENITE (LEFT) AND GABBRO (RIGHT)
C401-4, plane light, ×21. Note uralite rims around pyroxenes.

FIGURE 2.—GROUP IV PLAGIOCLASE CONTACT WITH A GROUP II PLAGIOCLASE GABBRO
C411-3, plane light, ×7. Mafic mineral of the gabbro is uralitized pyroxene.

FIGURE 3.—SCHLIERENIZED SCHIST XENOLITH IN GABBRO
T140-1, × nicols, ×7. Note chlorite bands (cl) along contact.

FIGURE 4.—TALC (t) DEVELOPING FROM TREMOLITE-ACTINOLITE (a)
Thistle Mills, plane light, ×59.

FIGURE 5.—CHLORITE BAND IN GABBRO
C334-1, plane light, ×59. Opaque mineral is magnetite.

FIGURE 6.—CHLORITE BAND IN QUARTZ DIORITE
T174-1, plane light, ×21. Band follows a joint.

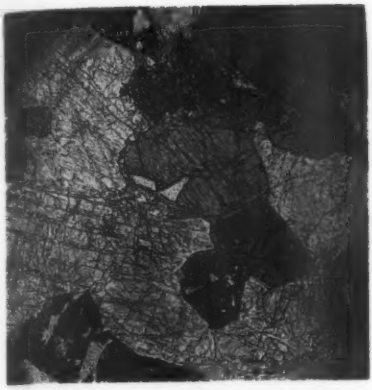


FIGURE 1

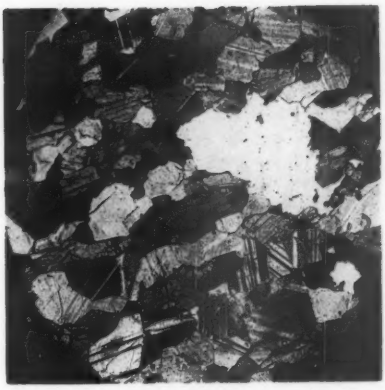
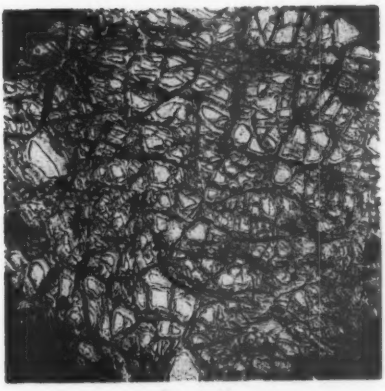


FIGURE 2

FIGURE 3



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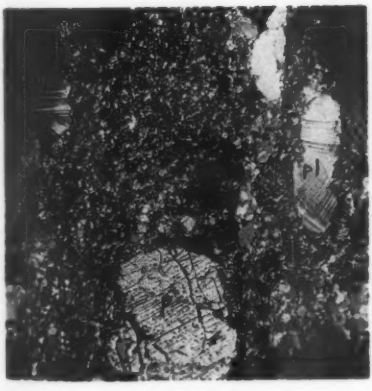


FIGURE 5

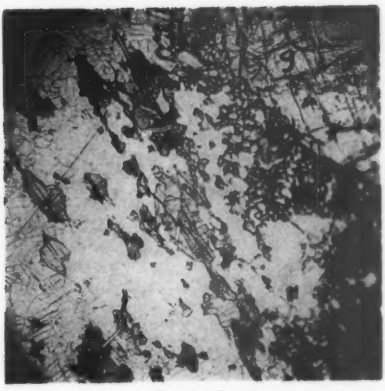


FIGURE 6

PHOTOMICROGRAPHS

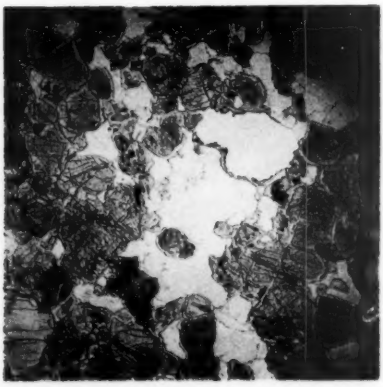
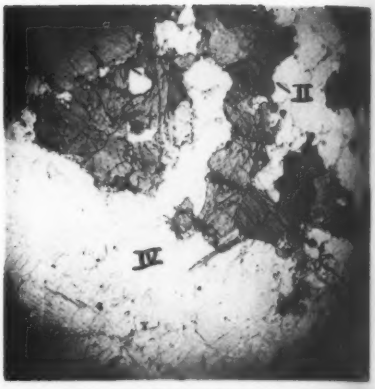


FIGURE 1



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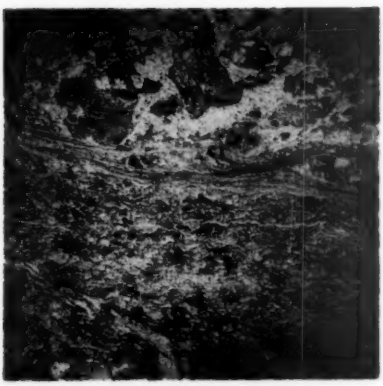

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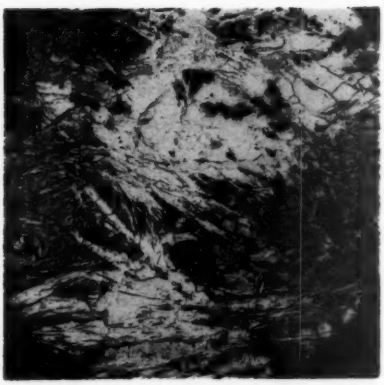

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FIGURE 6

PHOTOMICROGRAPHS

FIGURE 1

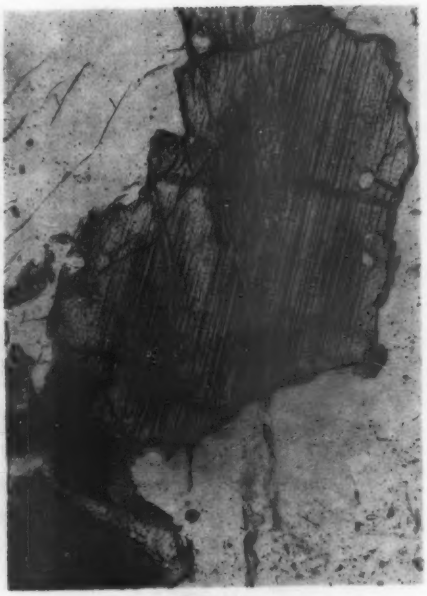

FIGURE 2

FIGURE 3

FIGURE 4

PHOTOMICROGRAPHS

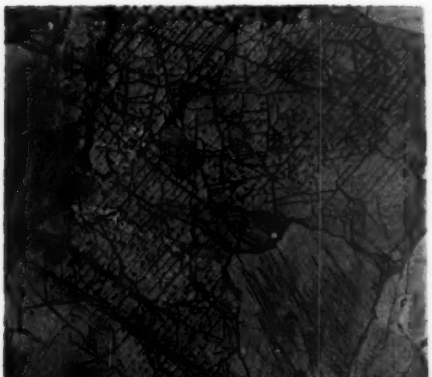

FIGURE 1

FIGURE 2

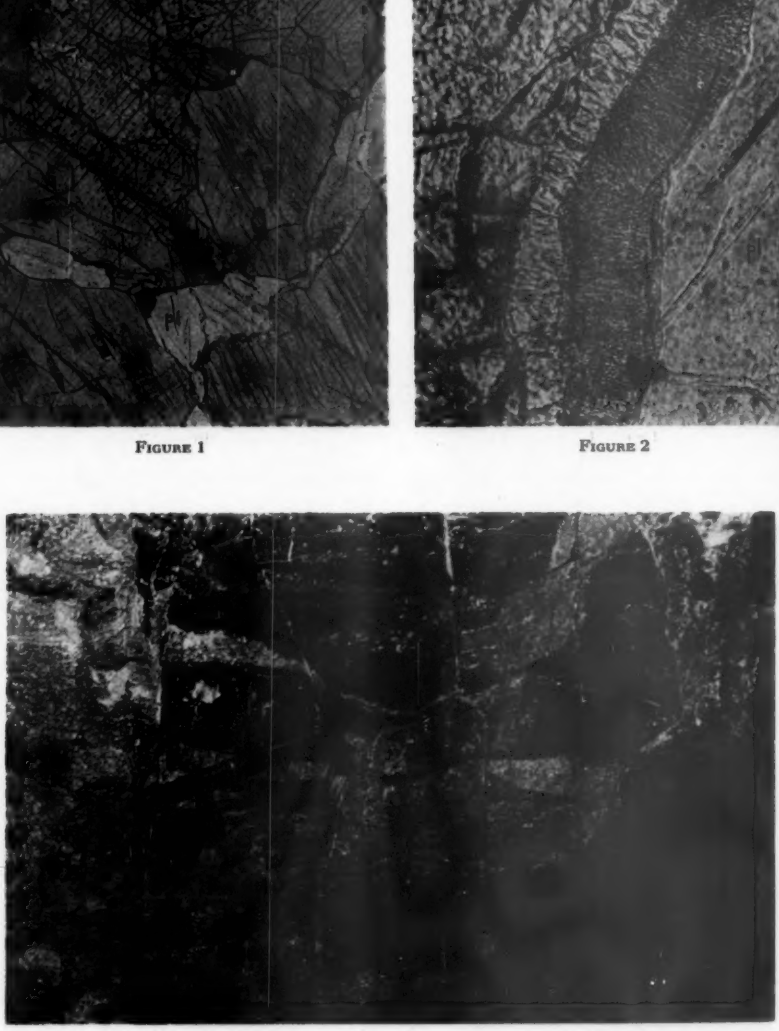

FIGURE 3

PHOTOMICROGRAPHS; GABBRO XENOLITH

about one half to 2 mm. It is usually colorless in thin section, but occasionally, as in Williams' slides, it is pale green with no pleochroism. Twinning, based on (100) as the composition face, was observed. Williams (1886) reported twinning inclined to the (110) cleavage varying from 25° to 35°. This may be the rare (101) twin (Winchell, 1933) since (110) (101) is about 31°. The presence of parting on both (100) and (001) makes the variety name diallage appropriate.

Plotted determinations of the clinopyroxenes before correction (Fig. 9) reveal 10 (38%) in the diopside field (nomenclature based on Hess, 1941), 11 (42%) in augite, 3 (12%) in endiopside, and 2 (8%) in salite. After correction, 17 (65%) are in augite and 9 (35%) in endiopside. The range in corrected determinations is $En_{46}Fs_5Wo_{29}$ to $En_{40}Fs_{21}Wo_{39}$.

Large zoned clinopyroxenes differing in optic angle and extinction from smaller, volumetrically dominant crystals were found (in C459-1 zoned crystals were 1.2 mm., rock otherwise 0.7 mm.) but their composition could not be accurately ascertained. An analog is drawn to the orthopyroxenes where early, zoned, magnesium-rich crystals were found coexisting with the normal more iron-rich type. This dominant group grades, with increasing ferrosilite molecule, from ultramagnesian rocks into gabbroic (Fig. 10).

TABLE 7.—OPTICAL DATA ON CLINOPYROXENES

Specimen	2V	Zᴀc	Beta	Estimated			Related		
				En	Fs	Wo	En	Fo	Ab
T86-1	+50° to +52°	+41° to +43°	1.675	55	6	39	90	90	
T80-1	+55° to +57°	+42° to +43°	1.676	53	7	40	87	89	
C459-1	+56° to +58°	+41° to +42°	1.681	51	8	41	84		34
C424-1	+56° to +58°	+39° to +40°	1.684	48	11	41	80		36
E4-2	+44° to +48°	+44° to +47°	1.683	49	11	40	78		19
C366-3	+52° to +54°	+46°	1.690	43	16	41	70		48

The clino- and orthopyroxenes exhibit a straight-line iron enrichment relationship (Fig.

PLATE 3. PHOTOMICROGRAPHS

FIGURE 1.—EARLIEST STAGES IN DEUTERIC ALTERATION OF PYROXENE TO URALITE

C352-4, plane light, ×26. Uralite forms complete rims around the pyroxene grains. Light mineral is plagioclase An_{70}.

FIGURE 2.—ADVANCING DEUTERIC ALTERATION OF THE PYROXENES

C366-3, plane light, ×82. Uralite now cuts through the pyroxene, enlarging along fractures and cleavages.

FIGURE 3.—COMPLETE ALTERATION OF THE PYROXENES

Mount Washington, Baltimore, plane light, ×26. Uralite appears as an irregular poikilitic mat. Evidence for former pyroxenes is in rough crystal outlines and a suggestion of pyroxene cleavage. Grains and irregular areas of epidote are occasionally developed between the plagioclase and uralite (e).

FIGURE 4.—PYROGENIC HORNBLENDE

C397-1, plane light, ×82. Crystals are well developed prisms giving the rock a directed texture.

PLATE 4. PHOTOMICROGRAPHS; GABBRO XENOLITH

FIGURE 1.—FELDSPATHIC PYROXENITE

Dogwood Road, Baltimore County, plane light, ×26. Pyroxene crystals are large and well developed. Plagioclase, An_{81} (pl) is entirely interstitial to the pyroxenes. Note oriented inclusions in the orthopyroxene.

FIGURE 2.—REACTION RIMS BETWEEN PLAGIOCLASE (pl) AND OLIVINE (ol) IN OLIVINE GABBRO

T105-1, plane light, ×384. Orthopyroxene is the innermost zone to the olivine; an amphibole-spinel symplectic is the wider, fibrous outer zone.

FIGURE 3.—JOINTED GABBRO XENOLITH IN GRANITE

T 171+80.

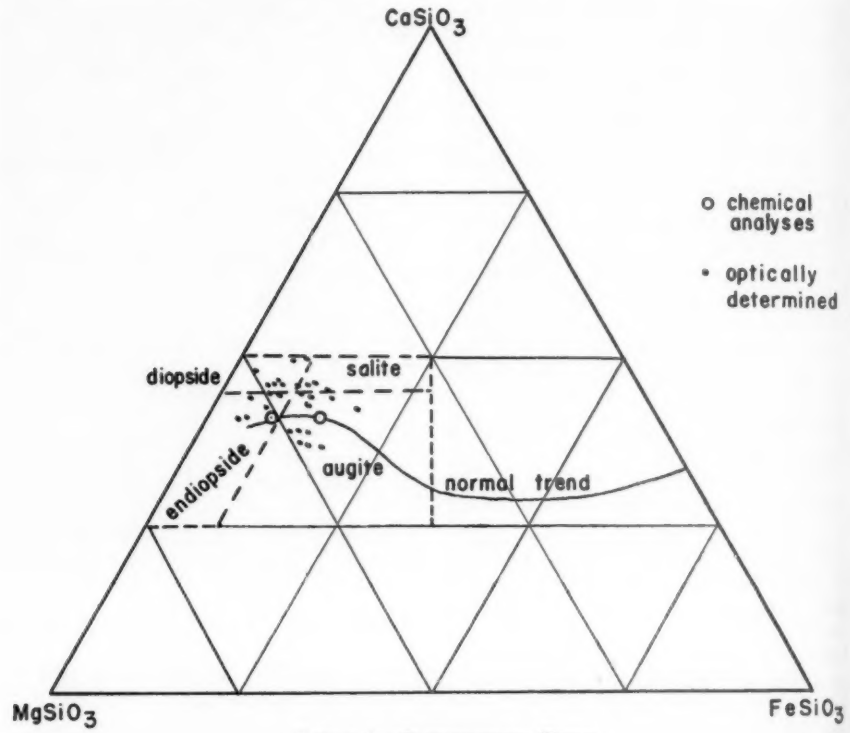

FIGURE 9.—CLINOPYROXENE TREND
Modified, Hess (1941).

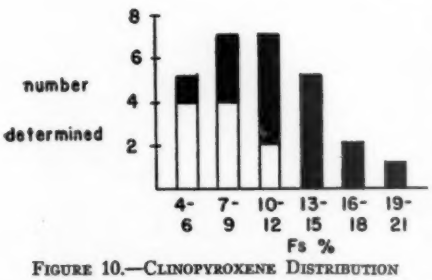

FIGURE 10.—CLINOPYROXENE DISTRIBUTION
27 determinations. Black = gabbros; white = ultramagnesics.

11). The clinopyroxene-olivine relation must therefore be similar to the orthopyroxene-olivine (Fig. 4).

Data for E4-2 can be taken as typical for determinations falling below the normal trend curve, i.e., with a 2V below 50° (Fig. 9). Five determinations in this area were in rocks with plagioclase An_{78-85} and the sixth An_{73}. The proximity of these points and their similar plagioclase seems to indicate a common origin and chemical composition, and perhaps the same minor constituent throwing them all off the normal curve to the same extent.

Inclusions are comparatively rare. Rods and plates as those in the orthopyroxenes are seldom encountered; possible exsolution of orthopyroxene material may exist, but could not be positively demonstrated. Reactions rims were observed, as with the orthopyroxenes.

Clinopyroxenes are easily altered by both serpentinization and uralitization. Serpentinization proceeds inwards along cleavage planes, usually as a fibrous antigorite. There was, however, no consistent optical relationship between the original crystal and the serpentine.

Uralitization preceded serpentinization, but also without any consistent optical relationship.

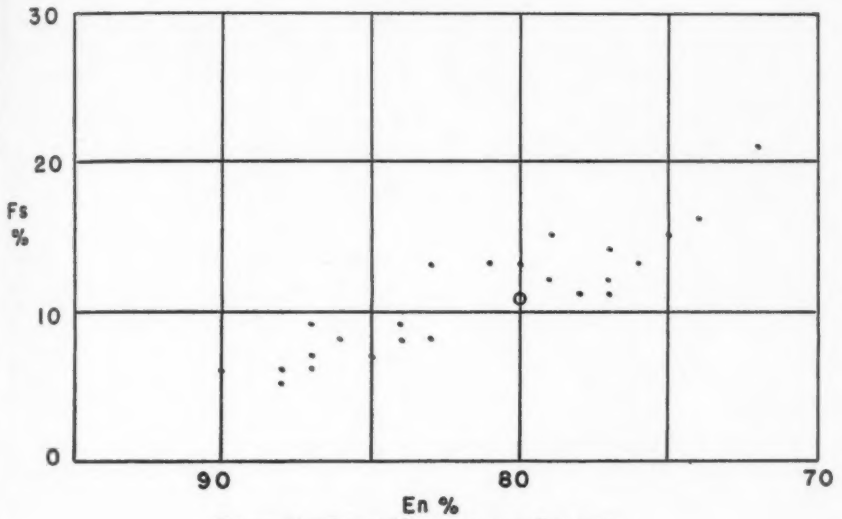

FIGURE 11.—ORTHO-/CLINOPYROXENE RELATION

In advances from the mineral contact inward, first with a greenish amphibole zone of fibrous or compact hornblende. With further alteration, the fibrous hornblende penetrates the clinopyroxene along cleavages. The innermost tremolite zone of the orthopyroxenes is invariably absent, probably because of higher alumina in the clinopyroxenes. (*Compare* tables 6 and 8.)

Amphibole

The optic angle, Z∧c, and the beta index were determined in every case. Exact compositions from optical data are impossible to obtain, but approximate compositions were figured from the information in an article by Sundius (1946) (Table 11).

The amphiboles are the minerals varying most in appearance. In uralitized varieties, they are invariably fibrous and anhedral. In most cases, the amphiboles and source pyroxenes do not develop a complete pseudomorphic relationship—smaragdite, a pale-green tremolite from diopside is a notable exception. The uralite usually starts as a fibrous rim about the pyroxene, in initial stages of development. In different slides, it is seen working its way inwards, sometimes along cleavages; other times, it has the appearance of an irregular exsolution along subparallel planes. The end product of this alteration is always a poikilitic crystal (Pl. 3, figs. 1–3).

TABLE 8.—CHEMICAL ANALYSES OF CLINOPYROXENES

	Clinopyroxene from Uralitized gabbro	Clinopyroxene from pyroxenite
SiO_2	51.41	51.80
Al_2O_3	4.32	2.21
FeO	9.31	3.50
Fe_2O_3		1.29
CaO	20.60	20.99
MgO	15.14	17.76
H_2O	nd	0.65
TiO_2	nd	0.13
Cr_2O_3	nd	0.51
MnO	nd	tr
	98.84	100.82
Specific gravity	3.35	3.308
Formula	$En_{43.2}Fs_{14.7}Wo_{42.1}$	$En_{50.2}Fs_{7.4}Wo_{42.4}$
Name	augite	endiopside
Analyst	W. G. Brown	T. M. Chatard

When the amphiboles are primary, they are subhedral or euhedral and give a well defined directed texture to the rock. They are apparently synchronous in developing with plagioclase in this rock type and never show any

tendency towards a poikilitic development as they do in the uralities (Pl. 3, fig. 4).

Grain size in the hypidiomorphic amphibolite-type rocks is of the order of magnitude of three-fourths mm. by one third and larger; in uralitized varieties, it varys as the stage of uralitization and the original size of the pyroxene, up to 2 mm. In pegmatites, where it is usually the only abundant mafic, it may be 10 mm. and larger. Pleochroism varies from nil in magnesian tremolite and anthophyllite to light shades of yellow and brown in smaragdite, and darker shades of blue, green, and brown in hornblende and pargasite.

Compositions vary depending on mode of origin. In uralitized varieties, all amphiboles types occur and indicate a close relationship to their source pyroxenes (Table 9). Amphiboles not derived from pyroxenes alone are generally richer in iron (Table 10), as would be expected in a later pyrogenic mafic mineral.

TABLE 9.—SOURCE OF URALITE

Specimen	Amphibole Mg content	Ortho-pyroxene Mg content	Clinopy-roxene Mg:Fe ratio
T49-1....	hornblende—80%	81%	47:13
C352-4...	" 70%	78%	47:11
E2A.....	" 70%	76%	46:13
P4-4....	tremolite 100% anthophyllite	>90	

TABLE 10.—AMPHIBOLES AND RELATED ROCKS

Rock type	Amphibole Mg range	Associated Ab range
ultramagnesian.....	70–100%	—
gabbro.....	70–80%	17–55
bojite.....	60–80%	14–56
xenoliths and pegmatites...	50–90%	34–43
quartz diorites.....	50–65%	72–88

TABLE 11.—OPTICAL DATA ON AMPHIBOLES

Specimen	Pleochroism	2V	Z∧c	Beta Index	Mg / Mg + Fe	Related En	Related Ab
Tremolite-actinolite							
P4-5	colorless	−80°	13°–16°	1.614	100%		
T48-2	colorless	−86°	18°	1.620	90%		
T153-1	pale green	−79° to −83°	17° to 20°	1.637	75%		24
T55-3	pale green	−80°	13°–14°	1.642	65%		(sauss.)
Anthophyllite							
P4-4	colorless	−80°		1.620	90%		
Hornblende-pargasite							
T49-1	yellow-green	−80° to −81°	14°–15°	1.638	80%	81	17
T157-3	green	−72° to −78°	7°–22°	1.642	75%		
C411-3	green	−79°–86°	13°–15°	1.653	65%		34
T23-1	yellow-green	−71°–77°	18°–16°	1.663	60%		38
T215–2	blue-green	−65°–86°	13°–12°	1.672	50%		74
E2A	brown-yellow	+86°	13°	1.660	60%		

Amphibole generally is related to plagioclase (Table 10; Fig. 12) becoming iron rich with increasing Ab molecule.

Certain conclusions can be drawn from the chemical analyses (Table 12), despite one being inferior (no recorded water or other volatiles, adding only to 99 per cent) and the other incomplete. Amphibole in gabbro derived from pyroxene has about the same Mg:Fe ratio as the pyroxene and a much higher Al_2O_3. Amphibole derived from pyroxene of ultramafic

rocks may have a higher Al_2O_3 but not nearly as high as the amphibole in gabbro.

Williams (1886) distinguished a primary yellowish-brown hornblende from later colorless or green uralite. This brown hornblende often is rich in minute black inclusions, possibly magnetite, and always occurs in irregular patches without any external crystalline form. Further (p. 45):

> The compact brown hornblende, which exceptionally occurs as a constituent of the Baltimore gabbro, appears finally to succumb to the same changes which the pyroxene undergoes, although it resists the change much longer than this. It is frequently observed quite intact when the other bisilicates are completely altered, but has itself been seen in all stages of transition to the fibrous form.

These same observations were made on amphiboles of the Cortlandt series in New York (Williams, 1887): that the brown hornblendes were associated with pyroxene rocks, passing into them, and that green hornblendes were related to the mica-diorites.

Shand (1942) found that brown and green hornblendes occurred in both the diorites and norites of the Cortlandt and suspected the only reason for the color to be the degree of alteration. The only criterion for the separation of hornblende was crystal habit—prismatic indicating an early crystallization and poikilitic a later deuteric one.

Walker (1940) described brown and green uralite hornblende in the Palisades diabase.

TABLE 12.—CHEMICAL ANALYSES OF AMPHIBOLES

	Amphibole from uralitized gabbro	Amphibole from Smaragdite rock
SiO_2	49.24	49.16
Al_2O_3	9.01	5.63
$Fe_2O_3 + FeO$	10.58	10.62
MnO	0.04	nd
CaO	14.36	nd
MgO	15.84	nd
	99.07	
Specific gravity	3.09	nd
Pleochroism	green-colorless	pale green
Analyst	W. G. Brown	W. S. Bayley
Reference	Williams 1886	Williams 1886

Kunitz (1930) showed titania to be the probable cause of brown pleochroism in the basaltic hornblendes.

Brown poikilitic hornblende, then, is not necessarily primary because of its color, which may be due to alteration or the fortuitous presence of titania, but rather the result of later deuteric action, which was also responsible for the poikilitic green hornblende. The only valid criterion is external crystalline form; optical data cannot differentiate poikilitic from prismatic and brown from green. In the light of this interpretation, Williams' observation of brown hornblende in completely uralitized slides is easier to understand.

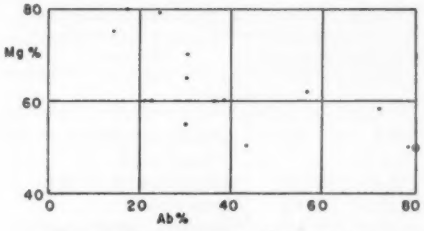

FIGURE 12.—AMPHIBOLE/PLAGIOCLASE RELATION

Opaque needle-like inclusions, never over 0.5 mm. long in the poikilitic hornblende are quite common. In prismatic sections, many of them parallel the c axis, or are inclined about 45°. In basal sections, they usually parallel the cleavage, though they may be random. The magnetic separator indicated them to be magnetite; the most magnetic fractions invariably contained hornblende with conclusions; hornblende in the other fractions would be clear. Rounded quartz grains occasionally occur as inclusions, usually with their longest dimension parallel the hornblende prism axis.

Amphibole, once formed, is stable. Thus, the end-stage deuteric alteration of much of the gabbro complex is a uralite gabbro or a serpentine-amphibolite. On rare occasions, in these latter rocks, amphibole was changed into talc, but with serpentine unaffected (Pl. 2, fig. 4).

Serpentine

Serpentine is common in all the utramafics (Table 13) and every olivine crystal was in some measure serpentinized.

Antigorite is the most common variety. No difference in optical properties was found between that derived from olivine (Pl. 1, fig. 4) and that from orthopyroxene (Pl. 1, fig. 3). Serpentine shows a fibrous structure in its early stages of development along olivine cleavages, and a later fibro-lamellar structure when this replacement is approaching completion. It may range from colorless to pale tints of yellow and green.

TABLE 13.—OPTICAL DATA ON SERPENTINE

Specimen	Beta Index	Occurrence
Antigorite		
T48-2....	1.580	gabbro pegmatite, from orthopyroxene
T75-1....	1.555	peridotite, from olivine
T131-1...	1.590	sheared zone
C457-4...	1.567	peridotite, from orthopyroxene
Chrysotile		
T32-1....	1.52	gabbro shear zone
T144-1...	1.511	ultramafic slickensides
C475-1...	1.535±	granulated gabbro zone

Chrysotile is less common and usually restricted to slickensided surfaces. Other varieties described (Johannsen 1928; and others) include picrolite, williamsite, deweylite, etc.

Clinozoisite

Clinozoisite is common in all the gabbroic rocks in which plagioclase has been saussuritized, often showing terminated crystals projecting into the plagioclase and more or less irregular edges into the amphibole. In bojites and quartz diorites, a panidiomorphic texture and unaltered plagioclase strongly suggests a primary origin.

Pistacite with a moderate amount of epidote molecule ($HCa_2Fe_3Si_3O_{13}$), typified by high birefrigence (±0.025), pleochroism, and higher index of refraction, was always associated with later intrusions and phases, and only occasionally with saussurite. Clinozoisite proper ($HCa_2Al_3Si_3O_{13}$) always occurred in saussurite (Table 14).

In uralitized rocks, clinozoisite usually develops along the contact between plagioclase and amphibole and forms continuous borders around the plagioclase, suggesting strongly that amphibole must contribute to its formation (Pl. 3, fig. 3). Chemical analysis reveals the ease with which clinozoisite may develop out of plagioclase with the addition of minor iron:

	Saussurite (soisite) from W692B
SiO_2	45.12
Al_2O_3	30.53
FeO	1.90
CaO	17.34
MgO	0.42
Na_2O	2.02
K_2O	1.09
H_2O	0.74
	99.16
Analyst	F. W. Clarke
Reference	Williams' notes

Accessory Minerals

Carbonates are common in minor amounts in altered rocks and in late joint-filling veins. Tremolite-actinolite rocks occasionally are altered to aggregates of talc-carbonate.

Laumontite is found as a well crystallized vein and cavity liner. Previous reported occurrences (Mather, 1937, p. 27) always put it near the quartz diorite contacts but in the tunnel, where preservation of this hydrated mineral was better, it occurs randomly.

Rarely, feldspar of ultramafic rocks altered to a zeolite mineral, appearing as a band of radiating tufts of colorless needles. Williams (1886) guessed this to be scolecite, which seems not improbable.

Other zeolites found at Woodberry and other quarries (Ostrander and Price, 1940) include leonhardite, stilbite, heulandite, apophyllite, natrolite, analcite, and chabazite.

Quartz is often observed in sheared zones, intrusive veins, and zenoliths (Pl. 2, fig. 3); occasionally in green amphiboles as slender elongate grains and in later rock types.

Tourmaline is associated with quartz-bearing felsic intrusions as prismatic crystals, nearly euhedral, and up to 6 mm. long.

Biotite and *muscovite* are associated with shear

zones, felsic intrusions, xenoliths, and later amphibolites.

Pyrite and other sulfides are widely distributed in insignificant amounts, often in equicracks and cleavages. It was not seen in unaltered ultramafics. Its occurrence in serpentine and its granular-massive habit is not inconsistent with an origin in serpentinization. Under

TABLE 14.—OPTICAL DATA ON CLINOZOISITE

Specimen	X∧c	2V	Y Index	Birefringence	Pleochroism	Occurrence
T153-1	+8°	+73°	1.710	anomalous blue	none	serpentine-amphibolite
C475-1	+5°	−85°	1.715	anomalous blue	none	uralitized gabbro
T2-1	0°+	−85°	1.734	+ 0.020	light yellow	bojite
T16-2	−7°	−78°	1.744	± 0.025	yellow	vein

TABLE 15.—OPTICAL DATA ON CHLORITE

Specimen	Beta Index	Other Data	Remarks
T0-1	1.58±	elongation−	clinochlore, plagioclase-magnetite rims in gabbro
H-1	1.604	optically+	penninite, garnet-amphibole rim in serpentine amphibolite
T14-1	(Z)1.626	optically+, birefringence .008, green	prochlorite in hydrothermal vein
T54-1	1.560	optically +, low 2V,	penninite in 1″ vein
T19-1	1.575	anomalous blue	penninite, associated with magnetite in gabbro
C336-2	1.580	birefringence	clinochlore, mixed schist-gabbro zone

granular rocks replacing both feldspar and pyroxene and up to three-fourths mm. in diameter. Crystals are usually deformed, except in quartz veins.

Apatite is frequently present in minor amounts as euhedral hexagonal prisms always associated with magnetite (Table 17). Dimensions vary considerably; in T14-2, it averages .07 by .4 mm.

Magnetite and other ore minerals are comparatively late. In some gabbros, magnetite forms a considerable portion of the rock, but it is generally insignificant (Table 17) averaging about one-fourth mm. In non-equigranular rocks, it gets up to 1 mm. In late chlorite veins, it may occur as stringers (Pl. 2, fig. 5); as a product of serpentinization, it may delimit former olivine crystals (Pl. 1, fig. 4).

Chromite is associated with the serpentine-amphibolites of the west and north borders of the complex. It is invariably associated with magnetite in a fine granular to massive habit and may also be seen in serpentinized olivine

TABLE 16.—DATA ON GARNETS

Specimen	Index	Specific Gravity	Occurrence
C506-2....	1.795	3.86	uralite gabbro, away from contacts
H-1.....	1.781	3.96	chlorite amphibolite at bojite contact (near schist)
C489-1....	1.802	3.88	hydrothermal vein
T135-2....	1.793	4.10	sedimentary xenolith

the microscope, the center of the grains appear opaque and the outside translucent red.

Chlorite is often found in fine veins traversing all rock types, averaging about .05 mm. in width, though getting up to several inches. It is also associated with hydrothermal alteration of the mafics, found especially in serpentine-amphibolites as a further alteration of the amphiboles. Plate 2, figure 5 shows the intimate relationship of chlorite to alteration, indicated by a feathery texture and apparent intergrowth into

mafic minerals. In the later rock types, the chlorite bands are clearly defined, following fractures, with well defined contacts (compare Pl. 2, figs. 5 and 6). It is also associated with hydrothermal alteration of the mafics, found especially in serpentine amphibolites as a further alteration of the amphiboles and forming kelyphitic borders around plagioclase and garnet (Table 15).

Talc occasionally occurs as an alteration product of tremolite or anthophyllite in serpentine amphibolites (Pl. 2, fig. 4).

Garnet occurs towards the Wissahickon schist contact. Elsewhere, a poikilitic garnet (C506-2) was found enclosing epidote, hornblende, and magnetite away from any contact; within intrusives in gabbro (C489-2); and in xenoliths of quartzite schist (T135-2). The material away from the contact was in poorly developed bands about 3 × 7 mm.; elsewhere, as dodecahedrons about 2 mm. in diameter. The largest found were up to 2 cm. in diameter, in altered ultramagnesian rocks, at Hollofield, generated by a bojite intrusion (Pl. 1, fig. 6), (H-1, Table 16).

Data for the xenolith, T135-2 (Table 16) are remarkably similar to data obtained from the Wissahickon schist itself and similar to those of the garnet developed in the gabbro near the schist contact.

Petrography

Introduction

Two broad rock groups are recognized: (1) Gabbroic rocks and their derivation products, and (2) Ultramagnesian rocks and their derivation products. They can be distinguished with ease in both thin sections and hand specimens, separated on the basis of plagioclase content; the former never below 40 per cent, and the latter rarely over 20 per cent.

Gabbroic rocks and their derivation products

The primary gabbros are holocrystalline, granular, medium-grained rocks, composed essentially of bytownite (usual range An_{78-82}) or a basic labradorite (usual range An_{64-70}) and ortho- and clino-pyroxene, with occasional minor olivine. Hornblende may proxy completely for the mafic minerals. Magnetite, sulfides, and apatite may or may not be present.

Uralitization and sometimes saussuritization are the alteration processes in these rocks. Uralitization of the pyroxenes leads to both colorless tremolite-actinolite and pleochroic green or brown hornblende.

The hornblende gabbro of previous authors, based on "primary brown hornblende," is not the hornblende gabbro or bojite and of this paper. It was previously believed that certain hornblende of the complex was primary, largely on the basis of its brownish color. Texturally, it was the same as the uralite hornlende, that is, poikilitic, and was often found with uralite rocks, even when all the other primary mafic minerals were gone. Cohen (1937) found this brown hornblende in distinctly foliated rocks, foliation always typical of uralite gabbro. He further found the brown hornblende to have the same refractive indices as the uralite green hornblende, and not basaltic hornblende, as might be expected. In different sections, other evidence, including chemical, has been introduced to show the green and brown hornblende identical.

It appears certain that this alleged primary brown poikilitic hornblende is merely another uralitization phenomenon and should be so called. In this paper, primary hornblende was recognized by texture; color was not considered any criterion. Prismatic hornblende, considered primary, usually occurred in rocks with little or no pyroxene, (Pl. 3, fig. 4). Brown or green poikilitic uralite hornblende considered secondary, was associated with a pseudomorphic change in pyroxene (Pl. 3, fig. 3).

Practically all the gabbros fall into Johannsen's (1939) family 2312P, gabbro and norite; occasionally in 3312P, melagabbro. Table 17 reveals several sub-groups distinguishable in thin sections:

(1) Olivine gabbro (T105-1) Johannsen 3312P—medium—grained (0.7–2.0 mm.), hypidiomorphic equant. Characterized by some olivine and plagioclase over 40%. Always shows reaction rims around plagioclase where it contacts olivine. Not too common.

(2) Norite (C445-1) Johannsen 2312P—medium—grained (0.7–1.5 mm.) hypidiomorphic equant. Reaction-rimmed orthopyroxene is only pyroxene. With accessory magnetite in west part of complex. Rare.

(3) Hyperite (TO-1) Johannsen 2312P—medium-grained (1.0–2.0 mm.) hypidiomorphic equant. Orthopyroxene equal to, or greater than, clinopyroxene. With accessory magnetite in west. Not uncommon.

(4) Bytownite gabbro (T49-1) Johannsen 2312P—medium—grained (1–4 mm.) hypidiomorphic equant, plagioclase tends to form the larger crystals. Characterized by plagioclase An_{78-88}, and clinopyroxene dominant over orthopyroxene. Magnetite may be present in minor amounts. Common.

(5) Labradorite gabbro (T19-1) Johannsen 2312P—identical to (4) except the plagioclase is An_{64-70}. Common.

(6) Magnetite gabbro (T14-2) Johannsen 2312P—identical to (4) or (5) except magnetite is about 5%, usually with apatite. Tends to be more plagioclase-rich than normal gabbros. Common, especially in the west.

TABLE 17.—SELECTED MODAL ANALYSES OF GABBROS AND ULTRAMAGNESIAN ROCKS

	T105-1 Olivine Gabbro	C445-1 Norite	T0-1 Hyperite	T49-1 Bytownite gabbro	T19-1 Labradorite gabbro	T14-2 Magnetite Gabbro	H-1 Bojite	T23-1 Uralite gabbro
Plagioclase	41.3	51.8	63.2	59.4	61.1	68.8	71.8	59.0
Clinopyroxene	41.2	—	16.6	29.7	29.1	16.8	—	—
Orthopyroxene	14.5	31.7	19.3	9.2	9.8	7.7	—	—
Olivine	3.0	—	—	—	—	—	—	—
Magnetite	—	5.	0.3	—	—	5.4	—	—
Apatite	—	—	—	—	—	2.0	—	—
Amphibole	—	11.3†	0.6†	1.7†	—	—	27.3*	38.1†
Clinozoisite	—	—	—	—	—	—	0.9	2.9
	W69 Feldspathic Lherzolite	T127-1 Lherzolites	T86-1	T8-1 Websterites	W386	W54 Altered Feldspathic Lherzolite	W174 Altered Lherzolite	P4-5 Smaragdite Rock
Plagioclase	17.6	0.6	—	0.6	—	7.3	—	—
Clinopyroxene	32.8	63.5	31.5	67.6	48.9	29.7	11.0	—
Orthopyroxene	42.8	13.7	58.8	28.1	51.1		36.1	—
Olivine	0.6	22.2	9.7	—	—	3.6	3.4	—
Magnetite & ores	4.8	—	—	—	—	7.7	10.0	1.7
Uralite	**	—	—	3.7	—	**	**	50.9
Serpentine	1.4	—	—	—	—	51.7	39.5	41.3††
Carbonate & talc	—	—	—	—	—	—	—	4.7
Clinozoisite	—	—	—	—	—	—	—	1.4

† Uralite
* Primary
** Very little, figured with serpentine
†† Includes chlorite

(7) Bojite (H-1) Johannsen 2312P—fine to medium — grained, directed hypidiomorphic equant (this slide ½–1½ mm.). Characterized by primary hornblende the only or dominant mafic. Often as pegmatitic-schlieren. Not uncommon, especially about contacts (see below).

(8) Uralite gabbro (T23-1) Johannsen 2312P—grain size dependent on source. Characterized by poikilitic uralite amphiboles. Most common type.

The term bojite, applied for the first time in this area, is proposed for several reasons. Weinschenk originally used it (1899) to denote amphibole gabbro with primary hornblende as its essential mafic mineral to distinguish it from

that containing uralite. Hornblende gabbro had been used previously here to denote a gabbro containing what is often a uralitized brown hornblende, and what should be called uralite gabbro. Cohen (1937) used amphibolite to denote bojite. He stated that it exhibited a crystalloblastic texture, but this author never observed it. Palimpsest structures, as pseudomorphic amphibole after pyroxene should have been present if the bojites were truly metamorphic, but are not seen (Pl. 3, fig. 4). The bojites are holocrystalline, hypidiomorphic, and many show not the slightest change. Plagioclase is completely free of any saussuritization, as often occurs in uralite gabbro, but is still clear, retaining its original crystal boundaries (Pl. 1, fig. 2). There seems little doubt that bojite represents a primary igneous rock type.

Rock terms used in this paper will include olivine gabbro, gabbro (norite, hyperite, bytownite-, labradorite-, and magnetite-gabbro) bojite, and uralite gabbro.

Ultramagnesian rocks and their derivation products

The ultramagnesian rocks are holocrystalline, granular, medium- to coarse-grained, composed essentially of orthopyroxene and clinopyroxene. Olivine may be present in large amounts or may be totally absent. Plagioclase, rarely over 20 per cent, is a bytownite, usually An_{78-88}; more often it is totally absent. Magnetite and chromite may or may not be present.

The alteration products include primarily serpentine and amphiboles, with varying amounts of chlorite, talc, magnetite, chromite, and carbonate. The serpentine is largely antigorite though some chrysotile has been noted. The amphiboles include tremolite-actinolite and hornblende, with minor amounts of pargasite and anthophyllite.

The ultramagnesian rocks fall into Johannsen's (1939) family 3312P melagabbro, 4112P websterite, and 418P saxonite. In thin sections (Table 17):

(1) Feldspathic lherzolite (W69) Johannsen 3312P — Hypidiomorphic inequigranular. Mafic grain size 0.7–1.5 mm.; plagioclase, 0.2–0.8 mm. Characterized by minor plagioclase (under 20%) with pyroxene inclusions, and interstitial to the phenocrystic pyroxenes. With no olivine, this rock is a feldspathic websterite. Common.

(2) Lherzolite (T127-1, T86-1) Johannsen 418P —hypidiomorphic equant, medium- to large-grained (1–8 mm. average 1.5–2 mm.). Characterized by a lack of plagioclase and the presence of olivine. The mafics may occur in any proportion. Common.

(3) Websterite (T8-1, W386) Johannsen 4112P—hypidiomorphic equant, medium- to large-grained (1–8 mm., average 1.5–2 mm.). Pyroxenes only mafic present. Not common.

(4) Serpentine rocks (W54, W174)—grain size dependent on source. Most show pyroxenes scattered through a serpentine matrix, with olivine remnants in the matrix, though pyroxenes may also be serpentinized. All degrees of alteration exist. Common.

(5) Serpentine- and chlorite-amphibolites (P4-5) —grain size variable. Characterized by amphibole in a serpentine or chlorite mat. All degrees of alteration are found. By further alteration of amphibole, more chlorite and talc appear. Most common ultramagnesian rock.

Williams called for transition rocks olivine-bronzite-gabbro with 25–30 per cent plagioclase between the gabbros and ultramagnesian rocks. Statistical study, however, reveals that olivine gabbros are generally lower in plagioclase than other gabbros, but never below 40 per cent. They differ from feldspathic peridotites, with plagioclase rarely over 20 per cent, in that they are equigranular. The peridotite feldspar is always interstitial and averages much smaller in grain size than do the mafics (Pl. 4, fig. 1). Texturally, then, there are no true transition rocks, and different processes seem to have operated in producing the gabbros and ultramagnesics.

Rock terms in this paper will include peridotites (lherzolite), pyroxenites (websterite; both with prefix feldspathic-); serpentine, and serpentine- and chlorite- amphibolite.

Structure

Cohen (1937) did the most detailed work on the structure of the gabbro complex. He mapped every discernable lineation and planar feature in search of a mode of emplacement.

Although his conclusions differ from those of this author, the more significant of his structures merit consideration.

Cohen described directed structures (pp. 225–227):

Banding—an alternation of tabular bodies varying only in the relative amounts of constituent minerals, with or without variation in grain size. Bands generally range from a fraction of an inch up to 2 inches or more and may wedge out or vary in composition until they grade into an adjacent band.

Schistosity in amphibolites (bojite of this paper)—due to a subparallel elongated or flattened folia of more or less aligned hornblende grains. The hornblende crystals are cleavage prisms with jagged terminations. Linear or planar schistosity may be developed; in the latter case, the prism axis c, and the b crystallographic axis lie in the same plane. Generally, banding and schistosity are parallel.

Cross joints in amphibolite—transect the linear schistosity at an angle of 90°. These are smooth, plane and subparallel.

Foliation in gabbro—pyroxene gabbro is massive; (brown uralite) hornblende gabbro is foliated, but is not important because of its rarity.

Fracture cleavage in serpentine—small subparallel fractures from an inch to several inches, and spaced, on the average about a fourth of an inch apart.

Schistosity in amphibole schists (chlorite amphibolite, etc. of this paper)—less developed than in the amphibolites; either a linear mineral arrangement or a planar cleavage.

Refracted joints—only south of Hollofield. Vertical in the less basic bands and less steep in the more basic ones.

Schistose zones—from 1 to 6 inches thick of intensely schistose amphibolite. Bulge, narrow, curve, and even fork. Some have an orientation and linear schistosity parallel the linear structure in the amphibolites and were probably formed at the same time.

Flexures of schistosity—only in places.

Quartz veins—numerous. Generally follow schistose and flexed zones.

Contacts—conformable and concordant.

Reasoning largely from these structural data, Cohen first concludes that the amphibolites have a metamorphic origin (p. 232–236):

1. amphibolites do not occur in dynamo-metamorphosed terrains.
2. their texture is crystalloblastic.
3. primary banding and later cleavage are not always parallel.
4. the primary banding—secondary cleavage relationship in the gabbro is paralleled in the wall rocks. (This is contradicted by the editor of Volume XIII, p. 236, stating that Broedel and Hershey have shown wall rock structures not parallel to those of the gabbro.)

The writer considers these rocks only as a result of primary differentiation. Invariable conformability between amphibolite foliation and the boundary of the gabbro mass strongly suggests primary intrusion; a metamorphic process causing the concentric formation of these rocks directed outward from the schist, without a concurrent formation of any sort of metamorphic zoning, is hard to picture. In the aqueduct (Herz, 1950), the bojites were seen intrusive into both the neighboring schist and ultramagnesian contact phases of the gabbro. They parallel the foliation of these older rocks, but, however, occasionally transgress them (hence Cohen's reason 3). Crystalloblastic texture is never discernable; alleged metamorphic mineralogic changes are more easily attributed to normal late magmatic conditions. Lastly, the usual interpretation of shear planes paralleling foliation seems most applicable here; that is, the rock crystallized under conditions of great deformation.

The flexured schistose zones show unaltered plagioclase, either with the same Ab:An ratio as the dominant gabbro or slightly more sodic (Pl. 2, fig. 2). Hornblende is usually larger than the pyroxenes in the contact area—which are often undeformed. These zones are then bojitic, a gabbro pegmatite phenomenon, due to later crystallization of wetter fractions (Pl. 6). This phenomenon, called liquation-fractionation by other authors, has been well documented, as from the Palisades of New Jersey (Walker, 1940) and the Whin Sill of England (Tomkieff, 1929).

Primary and so-called secondary structures are then best explained in terms of tectonic

conditions imposed upon a crystallizing or emplacing magma. Parallelism of schistosity and foliation become at once apparent as shearing in the direction of the forming foliation plane in a crystalline or semi-crystalline mass. The great mobility of the bojite schlieren certainly indicates a not yet completely solid rock.

Under this interpretation, the lack of agreement between structures within and without the gabbro body (Broedel, 1937; Hershey, 1937; Cloos, 1937) makes better sense. The prime mover in the gabbro was the crystallizing magma and its emplacement. Outside, as in the Wissahickon schist, metamorphism was responsible for structural development. Thus the schist must have already been metamorphosed when the gabbro was intruded. The gabbro emplacement indicates an adjustment to space already available, determined largely by existing gneiss domes, the Towson, Chattolanee, Woodstock, and Baltimore (Fig. 1).

Because of their essential parallelism, all planar features were plotted in one statistical diagram and linear features, including mineral orientations and slickensides, in another (Fig. 13). Data were obtained by the mapping of the Baltimore-Patapsco aqueduct. The city section shows the complex from its center to eastern margin, dipping west about 40° and striking northeast; the county section shows from the center to about 4000′ from the western margin dipping east steeply and striking northeast parallel the contacts. Well developed lineations point down dip, as expected in a magmatic development.

Interpretation of the city-section joints is difficult except for one minor concentration, a set of southeast-dipping cross joints. Since the gabbro is strongly intruded towards the southeast by quartz-diorite, it is not surprising that the joint pattern was upset by the superposition of newer fractures. In the county section, two sets of joints are well developed—one almost horizontal, actually dipping SSW about 10°, the other vertical and striking northwest. The first set is a true cross joint; the second is perpendicular to the first and the planar features and includes the lineations, and thus is a strike joint (Cloos, 1937) or an analogous longitudinal joint (Balk, 1937). Both are often mineralized by chlorite, carbonate, quartz, zeolite, albitite, and other hydrothermal veins.

The statistical data bear out the contention that the intrusion is more or less funnel-shaped. Though the southeast lip of the funnel dips less steeply than the northwest (not shown), the structure as a whole is a synclinal pluton. Further, the agreement of planar data, lineations, and joints to a primary igneous interpretation is significant.

Petrology

Comparisons with World Rock Types

Comparing the averages of chemical analyses of rock types of the Maryland-Pennsylvania area (Bascom, 1935) with the general averages for equivalent rocks of the world (Daly, 1933), reveals a consistently lower value for alkalies in the Maryland-Pennsylvania sequence (Table 20). This is emphasized by the highly calcic alkali-lime index (Peacock, 1931), 65. It is interesting to note that other typically calcic series, as the Lassen Peak (index 62.4) and Katmai (63.8), are also identified with orogenesis.

Differentiation Processes

Introduction.—Processes affecting the differentiation of the gabbro complex may be considered under two general headings: (1) crystal fractionation, and crystal settling—led to all primary differentiates of the complex; (2) deuteric and hydrothermal action—included the segregation of volatile-rich phases of the complex itself and the probable acquisition of volatiles from the wall rock.

The effects of differentiation and crystal settling in the magma were studied with a variation diagram. The orthodox diagram based on SiO_2 was found unsatisfactory, but a rectangular diagram based on iron enrichment, i.e., an abscissa of $\frac{FeO+Fe_2O_3}{MgO+FeO+Fe_2O_3}$ (*see* Wager and Deer, 1939, and Walker and Poldervaart, 1949, for further discussion) proved acceptable (Fig. 14). In it two stages are easily separable where the index is 31.1 to 36.0 and no analyses fall. The appearance of primary hornblende distinguishes a third phase about ratio 65, but it

FIGURE 13.—STATISTICAL DIAGRAMS

TABLE 18.—CHEMICAL ANALYSES OF ROCKS

	(1) Websterite	(2) Websterite	(3) Smaragdite Rock	(4) Websterite	(5) Websterite	(6) Feldspathic Pyroxenite	(7) Gabbro
SiO_2	53.98	52.55	53.22	50.80	53.21	48.91	45.35
Al_2O_3	1.32	2.71	3.14	3.40	1.94	8.81	16.11
Fe_2O_3	1.41	1.27	—	1.39	1.44	1.04	3.42
FeO	3.90	4.90	7.95	8.11	7.92	9.52	3.30
MgO	22.59	20.39	20.09	22.77	20.78	15.19	12.32
CaO	15.47	16.52	14.44	12.31	13.12	14.69	18.04
Na_2O	u	0.27	t	t	0.11	0.64	1.26
K_2O	u		t	t	0.07	0.10	
H_2O	0.92	1.09	0.98	0.52	1.01	0.59	
CO_2	u	—	—	—	0.10	—	
TiO_2	0.15	0.14	—	—	0.26	0.37	
Cr_2O_3	0.53	0.44	0.23	0.32	0.20	0.15	
MnO	0.21	0.24	0.11	0.17	0.22	0.16	
Cl	u		0.26	0.24		—	
P_2O_5	t	t	—	t	t	t	
SO_3			t	t			
NiO					0.03		
ZrO_2					t		
	100.48	100.52	100.42	100.03	100.47[b]	100.17[b]	100.00
Normative Minerals[c]							
quartz	0.24	—	—	—	—	—	—
orthoclase	—	—	—	—	0.56	0.56	—
albite	—	2.62	—	—	1.05	5.24	—
anorthite	3.06	4.73	8.62	9.17	4.45	20.85	38.64
nephelite	—	—	—	—	—	—	5.40
diopside	58.07	60.95	50.10	41.11	48.24	41.86	35.01
hypersthene	34.25	22.94	38.08	34.60	41.95	15.52	—
olivine	—	4.98	1.98	12.11	0.41	13.38	11.11
magnetite	2.09	1.86	—	2.09	2.09	1.62	2.55
chromite	0.67	0.67	0.22	0.45	—	—	—
ilmenite	0.30	0.30	—	—	0.46	0.76	—

	(8) Gabbro	(9) Uralite Gabbro	(10) Norite	(11) Uralite Gabbro	(12) Uralite Gabbro	(13) Gabbro	(14) Gabbro
SiO_2	44.76	46.85	48.02	48.02	46.68	46.85	45.41
Al_2O_3	18.82	20.02	20.01	17.50	17.12	19.72	23.05
Fe_2O_3	2.19	2.30	1.13	1.80	2.18	3.22	1.52
FeO	4.73	4.60	7.29	7.83	7.61	7.99	8.35
MgO	11.32	10.16	10.05	10.21	10.34	7.75	5.89
CaO	14.58	13.84	11.42	13.16	13.46	13.10	12.52
Na_2O	0.89	1.32	0.51	1.48	1.75	1.56	0.76
K_2O	0.11	t	0.05	t	t	0.09	0.32
H_2O	2.53	0.88	0.67	0.79	0.88	0.56	1.52
TiO_2	0.13	0.30	0.23				0.62
P_2O_5	—	t	t		t		0.13
MnO	0.15	t	0.18		t		0.09
Cr_2O_3	0.08		0.03				
CO_{22}			0.25				
NiO			0.01				
	100.29	100.27	99.98[d]	100.79	100.02	100.84	100.18

TABLE 18.—*Continued*

	(8) Gabbro	(9) Uralite Gabbro	(10) Norite	(11) Uralite Gabbro	(12) Uralite Gabbro	(13) Gabbro	(14) Gabbro
Normative Minerals							
quartz	—	—	0.66	—	—	—	—
orthoclase	0.56	—	0.56	—	—	0.56	1.67
albite	7.34	11.00	4.19	12.58	14.67	13.10	6.29
anorthite	46.98	48.65	51.98	41.14	38.92	46.43	54.41
diopside	20.11	15.90	3.59	19.56	22.43	15.14	2.29
hypersthene	3.03	9.85	35.78	11.63	1.06	10.56	25.34
olivine	16.04	10.05	—	12.47	18.86	9.82	1.00
magnetite	3.25	3.25	1.62	2.55	3.25	4.65	2.09
ilmenite	0.30	0.61	0.46	—	—	—	1.22
apatite	—	—	—	—	—	—	0.31

	(15) bojite	(16) bojite	(17) serpentine amphibolite	(18) serpentinized feldspathic lherzolite	(19) serpentinized peridotite	(20) serpentine
SiO_2	43.42	44.04	39.20	41.00	43.87	40.06
Al_2O_3	22.37	20.01	4.60	7.58	1.64	1.37
Fe_2O_3	0.81	4.22	3.45	5.99	8.94	3.02
FeO	9.25	8.61	6.15	4.63	2.60	3.43
MgO	5.75	5.01	31.65	23.59	27.32	39.02
CaO	13.34	11.68	3.23	10.08	6.29	—
Na_2O	1.24	1.24	0.42	0.52	0.50	—
K_2O	1.13	0.15	0.14	—		—
H_2O	1.63	2.01	9.88	4.73	8.72	12.10
TiO_2	1.25	2.24	0.52	—	0.12	—
P_2O_5	0.10	0.52	t	—	—	—
MnO	0.06	0.28	0.20	t	0.19	0.09
Cr_2O_3	t	—	0.41	—	0.44	0.20
CO_2	—	—	—	3.62	—	—
NiO	—	0.01	0.30	—	—	0.71
	100.35	100.42[a]	100.15	101.74	100.63	100.00
Normative Minerals						
quartz	—	1.20	—	—	—	—
orthoclase	6.67	1.11	—	—	—	—
albite	5.24	12.58	3.14	4.19	4.19	—
anorthite	51.99	47.26	10.84	18.63	2.22	—
diopside	11.04	5.91	1.93	24.57	10.50	—
hypersthene	17.17	18.72	20.05	5.36	54.50	32.65
olivine	—	—	46.69	32.11	2.24	53.00
magnetite	1.16	6.03	5.10	8.82	8.12	4.41
chromite	—	—	0.67	—	0.67	—
ilmenite	2.43	4.26	0.91	—	0.15	—
apatite	0.34	1.34	—	—	—	—
hematite	—	—	—	—	3.36	—
nepheline	2.84	—	—	—	—	1.43

[a] difference
[b] FeS_2 0.03
V_2O_3 0.03
[c] Many from Buscom, 1935

TABLE 18.—*Continued*

[d] FeS 0.11
V_2O_3 0.02
[e] ZrO_2 0.10
V_2O_3 0.05
FeS_2 0.25

1. Websterite, Hebbville, Baltimore Co.; Bronzite-diopside rock, W392; Analyst, J. E. Whitfield ref (Williams, 1890, p. 42)
2. Websterite, Hebbville, Baltimore Co.; Bronzite-diopside rock; Analyst, J. E. Whitfield (Williams, 1890, p. 42)
3. Smaragdite Rock, Dogwood Road, Baltimore Co. Altered pyroxenite; Analyst, J. E. Whitfield (Clarke and Hillebrand 1897, p. 83)
4. Websterite, Johnnycake Road, Baltimore Co. Bronzite-diallage rock, W386; Analyst, J. E. Whitfield (Williams, 1890, p. 41)
5. Websterite, Oakwood, Cecil Co. Bronzite-diallage rock; Analyst, W. F. Hillebrand (Leonard, 1901, p. 159)
6. Feldspathic Pyroxenite, Orange Grove, Baltimore Co. Williams: "olivine gabbro—unique" W69; Analyst, W. F. Hillebrand (Williams, 1895, p. 674)
7. Gabbro, Falls and Windsor Roads, Baltimore; may be Hyperite, W85; Analyst, W. S. Bayley (Williams, 1886, p. 37)
8. Gabbro, Wetheredville, Baltimore Co. similar to 7, W170; Analyst, W. F. Hillebrand (Williams, 1895, p. 673)
9. Uralite Gabbro, Pikesville, Baltimore Co.; Plagioclase Ab_1An_8, W64; Analyst, L. McKay (Williams 1886, p. 37)
10. Norite, Oak Grove, Cecil Co.; Bytownite and hypersthene, some magnetite; Analyst, W. G. Hillebrand (Leonard, 1901, p. 151)
11. Uralite Gabbro, Baltimore city and co.; Average of 19 powders; Analyst, L. McKay (Williams, 1886, p. 37)
12. Uralite Gabbro, Windsor and Forest Ave., Baltimore Co., W87; Analyst, L. McKay (Williams 1886, p. 37)
13. Gabbro, Baltimore city and co.; Average of 23 powders; Analyst, L. McKay (Williams 1886, p. 39)
14. Gabbro, Dublin, Harford Co.; Analyst, Penniman and Browne (Insley, 1919, suppl.)
15. Bojite, Ilchester, Howard Co.; Anorthite-hornblende rock, W333; Analyst, W. F. Hillebrand (Williams, 1895, p. 673)
16. Bojite, Rising Sun, Cecil Co.; Bytownite and hornblende; accessory quartz, magnetite, apatite and titanite; Analyst, W. F. Hillebrand (Leonard, 1901, p. 146)

Hydrothermally Altered Rocks

17. Serpentine-Amphibolite, Ellicott City, Howard Co.; Altered cortlandite, W317; Analyst, W. F. Hillebrand (Williams, 1895, p. 674)
18. Serpentinized Feldspathic Lherzolite, Sudbrook Pk., Balto. Co., W54; Analyst, W. F. Hillebrand (Williams, 1895, p. 674)
19. Serpentinized Peridotite, Johnnycake Road, Baltimore Co.; Altered porphyritic lherzolite, W174; Analyst, W. F. Hillebrand (Williams, 1895, p. 674)
20. Serpentine, Broad Brook, Harford Co. "Deep green translucent variety"; Analyst, F. A. Genth (Leonard, 1901, p. 164)

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may be overlapped by pyroxene gabbro crystallization.

Stage 1—Limiting Ratios 15–35.

Minerals separating

olivine Fo_{78-94}, concentrated within Fo_{86-90} (forsterite-chrysolite)
orthopyroxene En_{80-92}, concentrated within En_{88-90} (hypersthene-bronzite)
clinopyroxene $En_{56-46}Fs_{5-12}Wo_{39-41}$, concentrated within $En_{56-49}Fs_{6-11}Wo_{39-40}$ (endiopside-diallage augite)
plagioclase An_{78-94} (bytownite)
accessories—chromite, magnetite.

Petrographic types

pyroxenite, peridotite, feldspathic peridotite, and pyroxenite.

TABLE 19.—MODAL COMPOSITIONS OF SOME CHEMICALLY ANALYZED ROCKS

	1. Websterite	4. Websterite	6. Feldspathic Pyroxenite	7. Gabbro	17. Serpentine Amphibolite	18. Serpentinized Feldspathic Lherzolite	19. Serpentinized Peridotite
Plagioclase	—	—	17.6 (An_{79})	65.55 (An_6Ab_1)	—	7.3 (An_{83})	
Clinopyroxene	49.8	48.9	32.8	15.28	1.6		11.0
Orthopyroxene	50.2 (En_{85})	51.1 (En_{86})	42.8 (En_{81})	8.03	5.6 (En_{83})	29.7 (En_{88})	36.1 (En_{90})
Olivine	—	—	0.6 (Fo_{88})	—	11.7	3.6 (Fo_{87})	3.4 (Fo_{87})
Magnetite, etc.	—	—	4.8	11.43	15.2	7.7	10.0
Apatite	—	—	—	1.12	—	—	—
Uralite	—	—	1.4	—	38.7	—	—
Serpentine	—	—	—	—	27.2	51.7	39.5

TABLE 20.—AVERAGE CHEMICAL COMPOSITIONS COMPARED

	Peridotite		Pyroxenite		Olivine Gabbro		Gabbro		Quartz Diorite	
	Md & Pa av 2*	Daly #78 av 13	Md & Pa av 5†	Daly #83 av 12	Md & Pa av 2‡	Daly #56 av 17	Md & Pa av 7	Daly #55 av 24	Md. av 2	Daly #47 av 125
SiO_2	41.37	43.95	51.40	52.33	47.77	46.49	47.41	49.50	58.86	58.90
Al_2O_3	3.11	4.82	2.08	3.54	14.37	17.73	18.60	18.00	16.80	16.47
Fe_2O_3	6.15	2.20	2.96	2.61	1.67	3.66	2.00	2.80	2.75	2.89
FeO	4.36	6.34	7.33	5.19	7.05	6.17	8.62	5.80	5.97	4.04
MgO	29.38	36.81	23.05	23.92	12.64	8.86	6.59	6.62	3.34	3.57
CaO	4.74	3.57	9.47	10.29	14.23	11.48	11.23	10.64	7.94	6.14
Na_2O	0.46	0.63	0.06	0.43	0.98	2.16	1.70	2.82	1.97	3.46
K_2O	0.07	0.21	0.03	0.35	0.05	0.78	0.42	0.98	1.04	2.11
H_2O	9.27	1.08	2.48	1.03	0.73	1.04	1.55	1.60	1.83	1.27
TiO_2	0.32	0.10	0.20	0.10	0.33	1.17	1.20	0.84	1.02	0.76
P_2O_5	—	0.10	0.15	0.06	—	0.29	0.23	0.28	0.29	0.27
MnO	0.20	0.19	0.18	0.15	0.08	0.17	0.20	0.12	0.16	0.12

* Cr_2O_3 0.42
† Cr_2O_3 0.43
‡ Cr_2O_3 0.07

Critical ratios

30, the start of plagioclase crystallization in peridotite.

MINERALOGICAL EVIDENCE: During the earliest stages of crystallization, conditions of non-equilibrium and strong fractionation prevailed. In some areas, early olivine was resorbed and pyroxenes crystallized; elsewhere, it was fractionated while being only partially resorbed. Conditions everywhere favored the development and zoning of large pyroxenes. In some areas, growth of early olivine was entirely counteracted by complete resorption; elsewhere, early olivine was protected by reaction rims or by fractionation into another environment.

Clearly then, there were two early lines of descent: (1) peridotites with higher magnesia because of dominant olivine forsterite, and (2) pyroxenite with higher silica and lime from their bulk of pyroxene (Fig. 14). In general, however, even in the peridotites, pyroxenes tend to be large while olivine is smaller and rounded. As would be expected, peridotites are more widespread than pyroxenites, but zoned pyroxenes in both attest to the lack of equilibrium during this early history. This almost

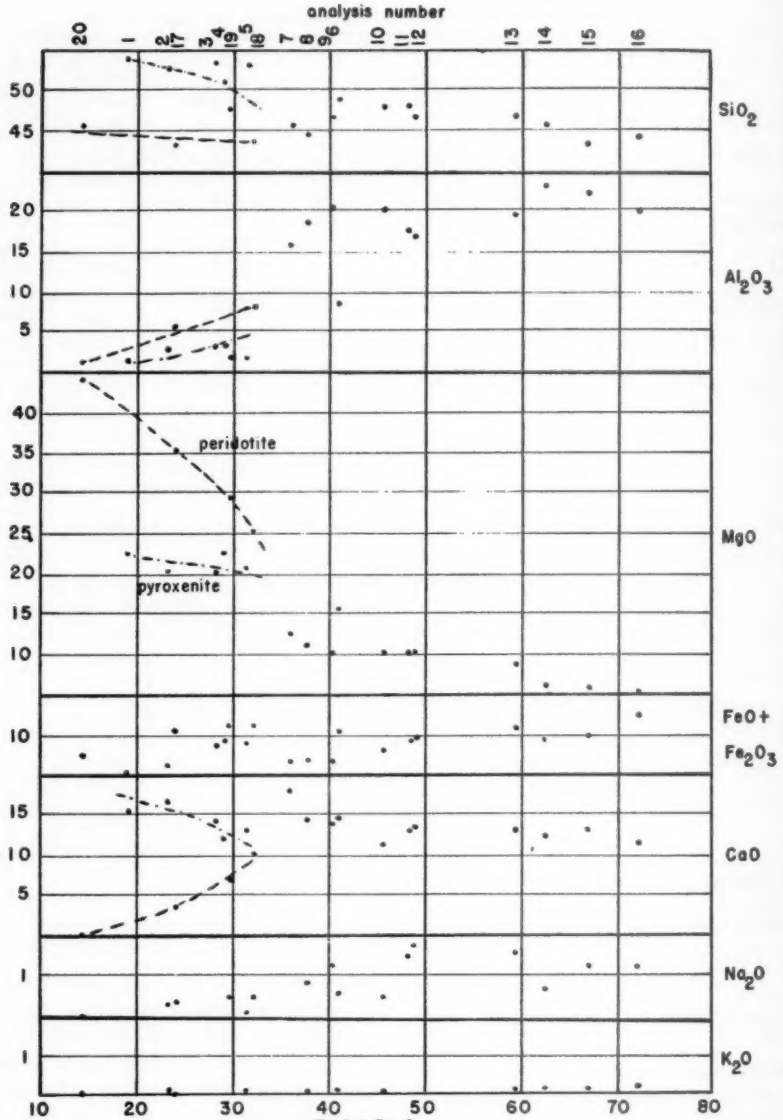

FIGURE 14.—VARIATION DIAGRAM OF GABBRO COMPLEX
20 analyses, recalculated water-free to a basis of 100 per cent.

purely mafic crystallization continued until plagioclase very suddenly and dramatically jointed the suite in very large volume.

Peridotite should be considered the more normal ultramafic because: (1) Mapping reveals it, together with its altered derivatives, to be

the most common ultramagnesian type. (2) All the pyrogenic minerals found in the volumetrically dominant gabbro, i.e., ortho- and clinopyroxene, plagioclase, and olivine, are present in this type, different only in magnitude. (3) Their oxide lines of descent trend directly into gabbro; pyroxenites do not match in silica, magnesia, and alumina (Fig. 14).

Whenever bytownite appears, it is minor in amount, interstitial to the mafic minerals, and severely strained. It is definitely late and must have developed while the pyroxenes were becoming zoned in a less magnesic environment after initial fractionation. The feldspathic ultramafics of this period certainly do not appear to be completely transitional to the gabbroic rocks, for the latter appear suddenly with plagioclase losing its interstitial nature for good and becoming equal to or greater than the mafics in volume.

OXIDE TRENDS: Points for peridotite, the more normal ultramafic, are from analyses that had significant water (up to 10 per cent and more), revealing a large measure of serpentinization. The analyses were recalculated water-free and to a basis of 100 per cent, then plotted. Nos. 17, 18, and 20, revealing significant olivine in their norms, give the best trend outline. Though discrepancies may be expected, the points indicate the trend from peridotite into feldspathic peridotite and finally into gabbro.

Silica reveals no consistent direction and leads directly into gabbro silica. This lack of variation is not surprising for the increase from anorthite (SiO_2=43.2%) replacing olivine is offset by a volumetric decrease in pyroxenes (enstatite SiO_2=60%). Alumina and lime increase as the clinopyroxenes become more important (*compare* tables 6 and 8), later, more steeply, by the appearance of plagioclase, largely as anorthite. Magnesia decreases rapidly because of the disappearance of olivine and the increasingly subsidiary role of orthopyroxene (En_{82}); its place taken by clinopyroxene (En_{52}). Iron increases slightly, going into the pyroxene molecules. Albite molecule accounts for the rise in soda from 0 per cent; potash is practically nonexistent.

Though the pyroxenite trends appear more subdued, about the points where their SiO_2, MgO, and CaO curves intersect the peridotite trends, ratio 35, gabbro appears. This seems to underscore the one line of descent when the liquid reached a certain composition and plagioclase joined the crystallizing suite in large volume.

Stage 2—Limiting Ratios 35–65.

Minerals separating

olivine Fo_{85-80} (chrysolite)
orthopyroxene En_{70-85}, concentrated within En_{77-90} (bronzite)
clinopyroxene $En_{51-40}Fs_{8-21}Wo_{41-39}$, concentrated within $En_{43-42}Fs_{11-16}Wo_{41}$ (augite)
plagioclase An_{64-86}, concentrated within An_{76-82} and An_{64-70} (bytownite-labradorite)
accessories—magnetite and apatite.

Petrographic types

olivine gabbro and gabbro.

Critical ratios

30–35, the beginnings of gabbro crystallization; under 45, the end of olivine; about 65, the start of primary hornblende.

MINERALOGICAL EVIDENCE: Zoning shows that fractionation of the pyroxenes was still going on. However, a true gabbroic environment had been brought about as indicated by an equigranularity of plagioclase and pyroxene. Olivine, which soon disappears, is always smaller than the other constituents and must have been strongly resorbed forming more pyroxene. Lack of plagioclase zoning indicates that it was not fractionated as strongly as pyroxene. Bent twinning lamellae, incomplete or exsolution lamellae, and granulated areas in both minerals, however, together with other structural evidence, indicates that syncrystalline deformation must have been quite strong, especially during the emplacement of the gabbroic rocks.

Analysis #6 (Table 19), called "olivine gabbro" by Williams, is described in his specimen list as "unique". It is just that, being more a feldspathic pyroxenite than anything else (only 0.6 per cent modal olivine). Its zoned and reaction-rimmed pyroxenes are larger averaging 1 x .5–.7 mm., than its intersertal plagioclase and olivine, and approach a pris-

matic form, as in the true pyroxenites (Pl. 4, fig. 1). This type of rock must have led Cohen (1937) and others to the conclusion that the pyroxenes were earlier than olivine; certainly this is not true in the more abundant peridotites. The early magnesic orthopyroxene, En_{91}, and the comparatively later olivine, Fo_{88} add further evidence to an unusual origin. This discrepancy was never found in true peridotites or gabbros.

Feldspathic pyroxenites originated in fractionation and incomplete crystal settling. Pyroxenes formed under conditions postulated for the pyroxenites, but, either because of a high magmatic viscosity or turbulence due to emplacement, ended up in a typically gabbroic environment. There strong reaction rims and zoning were developed while plagioclase and olivine crystallized out. Petrogenetically they are interesting, but their rarity rules out any generalizations towards a unique origin for the complex as a whole.

INTERPRETATION OF OXIDE TRENDS: Silica is nearly level. Alumina is very erratic, possibly increasing because of increased plagioclase and substitution for silica in clinopyroxene. Magnesia decreases while total iron increases; a complementary reaction from diminishing enstatite and increasing ferrosilite in the pyroxenes. Magnetite also increases, though always relatively minor. Lime and soda are close to level, indicating a lack of variation in plagioclase. This agrees with the plagioclase statistical study (Table 1). Potash is insignificant.

Stage 3—Limiting Ratios 65–80.

Minerals separating

hornblende
plagioclase An_{84-86}, concentrated within An_{78-83} and An_{64-70} (bytownite-labradorite)
accessories — pyroxenes, magnetite, apatite, and clinozoisite

Petrographic type

bojite

Critical ratio

65, the start of primary hornblende crystallization

MINERALOGICAL EVIDENCE: The bojites were the last pyrogenic phase and due to accumulation of volatiles (for other conditions see below). They often occur as pegmatitic facies and intrude both ultramafics and schist along the contacts of the massif, where volatiles were accessible from the country rock.

Evidence for a late age is indicated in other ways. The bojite hornblende is more iron-rich than hornblende derived from pyroxenes (Table 10), indicating an ability to crystallize less refractory iron oxides. Also, bojite often occurs as a micro-pegmatite schlieren filling cracks in gabbro (Pl. 1, fig. 5). Last, as felsic or pegmatitic bands in gabbro, it must have come as a liquid mesh in an already crystalline rock (Pl. 6).

The sudden change in primary mafic from pyroxene to hornblende has been often observed and described in some detail by Bowen (1928), Tomkieff (1929) and others. All state water to be one of the driving forces involved. That water was important here is indicated by the role bojite plays as a pegmatitic and micropegmatitic schlieren. Both the development of large crystals and great mobility are possible only in a volatile-rich medium.

Bowen (1928) found magma series crystallizing hornblende to have a general increase in alumina up to a maximum of about 18 per cent; at that point, hornblende started to crystallize. The alumina increase was considered due to a plagioclase-melt reaction producing sodic plagioclase, resulting in a less net substraction of alumina than if calcic plagioclase had to start crystallizing. These conditions are met in a large crystallizing body where differentiation goes on under slow cooling and limited fractionation.

This theory seems to fit the variation curve very well (Fig. 14). Alumina trends upward to 23.05 per cent at analysis #14, the last pyroxene gabbro. The bojites following indicate a downward trend and the start of hornblende crystallization. Plagioclase changing from An_{78-83} to An_{64-70} completes the picture. The environment change causing hornblende rather than pyroxene to crystallize was as dramatic and sudden as that causing plagioclase. Both came out in floods and without any extended series of transition rocks.

INTERPRETATION OF OXIDE TRENDS: Though only two analyses of bojites were available, gabbro trends lead directly into them without any suggestion of a break.

gaard intrusion trends, showing iron enrichment for 95 per cent of the crystallization, are plotted.

During the entire crystallization of the gab-

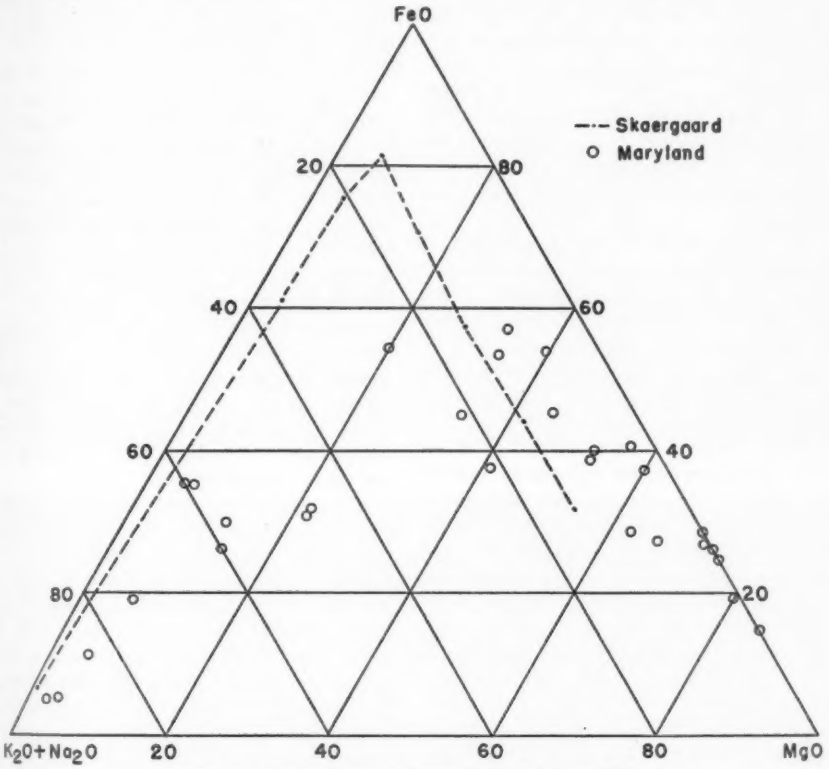

FIGURE 15.—DIFFERENTIATION TREND OF MARYLAND ROCKS

Differentiation Trends

Most petrologists today agree that a normal basaltic magma in differentiating will show absolute enrichment in iron for the major part of the crystallization period, followed in the last stages by enrichment in alkalies and silica (Wager and Deer, 1939; Walker and Poldervaart, 1949). The trends of the area were studied on a triangular diagram with MgO, FeO, and (Na_2O+K_2O) as apices (Fig. 15). Ultramafics and gabbros used are listed in Table 18. Non-gabbroic analyses were obtained from Williams (1895), Grimsley (1894), and Bascom (1902). For comparison, the Skaergaard intrusion trends, showing iron enrichment for 95 per cent of the crystallization, are plotted.

bro complex, strong ferro-magnesian fractionation, as reflected in the parallelism of the trend to the MgO-FeO side of the triangle, took place. Thin-sections showed further evidence with well developed mafic zoning; plagioclase zoning almost totally absent. Bojites are at the iron-rich end of the gabbro-ultramagnesian trend and mark the beginnings of strong felsic fractionation and the start of quartz bearing types. Since ferromagnesian fractionation did not go on as long in the gabbro complex as it did in the Skaergaard intrusion, for example, a fuller suite of quartz-bearing rocks developed in Maryland.

Banding

Alternation of gabbroic and ultramafic layers.—Away from the contacts, where ultramafics dominate, a banding between the two rock types occurs. The bands are highly variable, ranging from a few inches up to tens of feet. Variations in grain size may or may not occur, but ultramafic pyroxenes tend to be phenocrystic, much as they are in the larger ultramafic bodies; indeed the two textures are identical. There seems every evidence that both the bands and the larger bodies are identical.

It is easy to see why past workers considered these bands intrusive; compared to the allegedly intruded gabbro, they are tabular and narrow. However, there are no contact effects between the bands and the "intruded" rocks; indeed, study often reveals a gradational contact (Pl. 2, fig. 1). The main ultramafic masses, on the other hand, were emplaced before the major gabbroic intrusion. This is evinced at Hollofield where bojite intrudes a serpentine-amphibolite (Pl. 1, fig. 6) and generates strong contact effects in the ultramafic as chloritization for a thickness of about 20 mm., and develops large garnets. Similar observations were made in the aqueduct. Numerous autoliths of serpentinized and amphibolitized material within the gabbro offer final evidence, as in the transition zone of gabbro-ultramagnesics at the Patapsco River and U. S. #40.

Deformation in the crystallizing mass was probably the chief mechanism for the formation of these bands. Thus, Bowen states (1928, p. 168):

> The action is of the nature of the intrusion of the more completely liquid portions into rifts in the crystal mesh of the part of the mass in which the proportion of crystals is much greater. The process is thus sort of auto-intrusion.

Since the complex crystallized under deformation conditions, further credence is added to this view.

Variations of ultramafic layers within themselves.—Ultramagnesian rocks almost never occur massive. A banding is found within them due to minor changes in mineral content. This is best seen in altered varieties, as at Hollofield, where over 32 tabular bands were distinguished in a vertical distance of 117′ (Pl. 5). These bands never intersect and are continuous except for the smallest, as far as can be traced. Their thickness, at Hollofield, varies from 1½″ up to 14′ 7″; 12 of the bands measure under 1′; the overall average is 3′ 8″. Three of the bands are bojite causing contact effects; the others result from changes of composition in amphibolites (largely tremolite and anthophyllite), chlorite-amphibolites, serpentine-amphibolites, and talcose amphibolites.

An injection origin for the bands is impossible except for the bojites, because of their persistence, lack of xenoliths, and lack of contact effects. The banding appears rhythmically repeated and should be ascribed to gravity layering, perhaps combined with early auto-intrusion and further crystal sorting.

Alternation of more felsic material within the gabbro.—Highly irregular bands or schlieren of plagioclase or a pegmatitic gabbro often cut through the gabbro (Pl. 6). The plagioclase is generally slightly more sodic than that of the enclosing gabbro. Typically, C411-3 (Pl. 2, fig. 2) shows gabbro plagioclase An_{76} and An_{68} in the felsic band. Contact effects are lacking; mafic minerals in these bands are nearly always amphiboles and most often, hornblende.

The origin of these bands should be sought in volatile-rich fractions of the gabbroic magma where minerals of larger grain size could crystallize. Further, these fractions are perfect for the development of hornblende rather than pyroxenes, as has been pointed out. The formation of bojite elsewhere may have been slightly earlier or synchronous with the crystallization and emplacement of these bands.

Chloritic bands in all rock types.—Chlorite bands are very common in the ultramagnesics, and are also seen in the gabbros, ranging from

PLATE 5. BANDED ULTRAMAFIC ROCKS

FIGURE 1.—HOLLOFIELD QUARRY, BALTIMORE COUNTY
Banding seen is due to changes in secondary minerals.
FIGURE 2.—GWYNN'S FALLS NEAR GWYNN OAKS AVENUE, BALTIMORE
Note the parallelism of the bands.

FIGURE 1

FIGURE 2

BANDED ULTRAMAFIC ROCKS

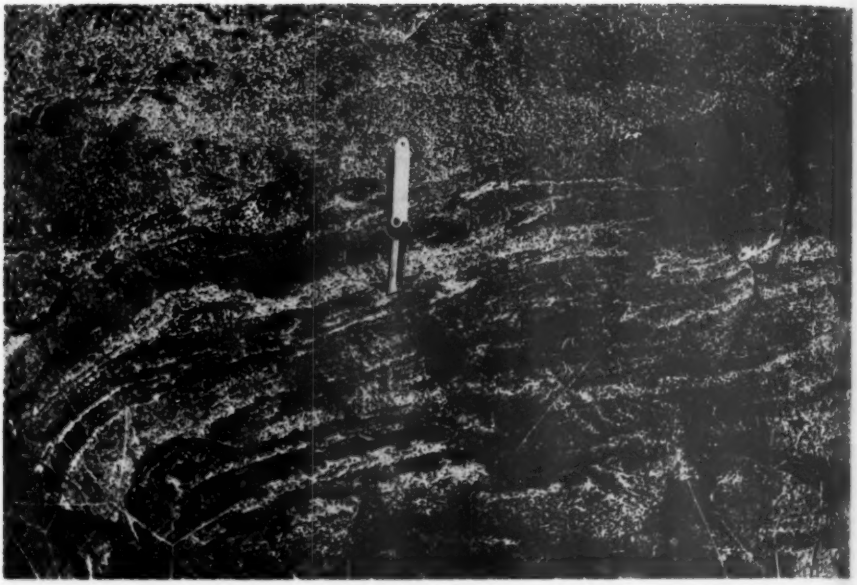

FIGURE 1

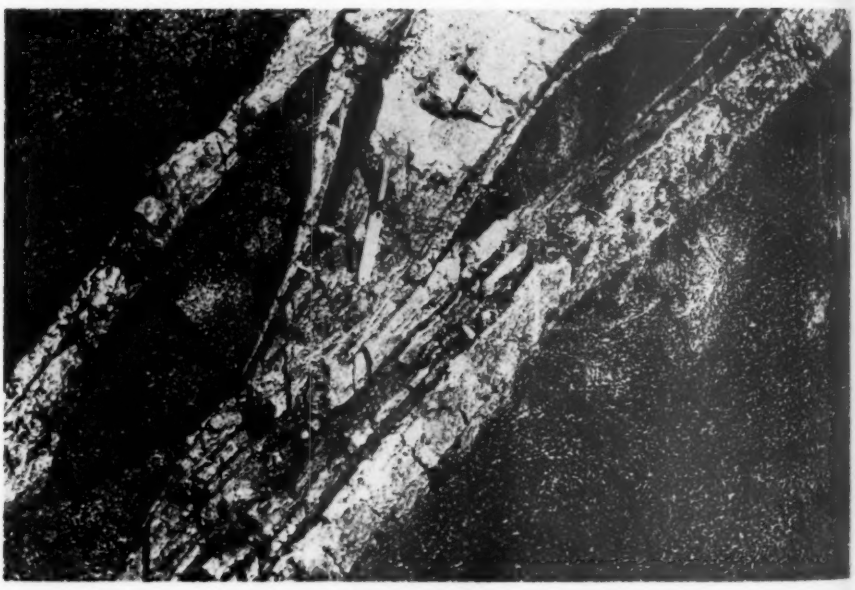

FIGURE 2

FELSIC AND BOJITIC SCHLIEREN

less than a millimeter to 2 inches or so. As a general rule, they do not follow any foliation but rather later fractures. Contact effects are always nil. This banding may be attributed to a late hydrothermal alteration of the mafic minerals along these joints (Pl. 2, fig. 5 strongly suggests such a process), or to the alteration of late hornblendic veins (Pl. 2, fig. 6).

Mechanism of Intrusion

All exposed sides of the intrusion reveal a concentration of ultramafic rocks. Since contact effects are not great, differentiation by gravity settling of early formed pyroxenite and peridotite must have taken place at depth, and the sides of the complex must have been crystalline at the time of emplacement, certainly by the time of the gabbro rest magma intrusion, as has been pointed out.

The development of the complex as a whole should be attributed to an auto-intrusion with a time interval between ultramafics and gabbroic rocks. The time of emplacement was one of great orogeny; the forces allowing ultramafic intrusion were later renewed to allow gabbroic emplacement, both within a downbuckle of Wissahickon schist. Syntectonic development of the massif is seen everywhere in the development of shear planes parallel the primary foliation. The complex as a whole seems to lie close to an axis of intense metamorphism, i.e., metamorphism decreases away from it towards the west.

The acts of intrusion were probably closer in time than Taconic and Appalachian as might be guessed. The lack of metamorphic facies development in the ultramafics and the gradational contact between them and the gabbro in places leads to the conclusion that the ultramafics heralded the beginnings of Appalachian orogeny; that gabbroic and later alkaline intrusives developed as the orogeny progressed.

Post-Magmatic and Hydrothermal Effects

The most important post-magmatic effects are the uralitization of pyroxene to either a colorless amphibole or hornblende and some saussuritization of the plagioclase. Hydrothermal effects include serpentinization of olivine and to a lesser degree pyroxene, and the development of talc and chlorite from uralite.

The amphibole of the altered ultramafics is generally actinolite or anthophyllite, a direct reflection of low alumina content. Hornblende occurs largely in the gabbros, due to the presence of plagioclase; further alteration of the amphiboles may form chlorite and talc.

The recent work of Bowen and Tuttle (1949) aids in an interpretation of the above changes. Considering their work and evidence already presented, the following picture seems not at all unreasonable. During the initial emplacement, some magmatic water was present. With falling temperature, the water-rich areas were uralitized. After this change, the temperature was raised, perhaps because of the intrusion of the gabbroic rocks, and serpentinization took place, near-by schists supplying most of the water this time. Olivines were first to be serpentinized; pyroxenes and their amphibole derivatives were more resistant to this change. Direct evidence of the lateness of serpentinization is that it took place after the development of reaction rims between plagioclase and olivine. The Bowen and Tuttle experiments revealed that, at low temperatures, olivine was more readily serpentinized than enstatite; the reverse was true at high temperatures. This adds credence to serpentinization occurring with rising temperatures. Furthermore, talc develops from amphibole and not directly from enstatite; if all changes resulted from cooling under wet conditions, a mass of talc serpentine would have developed directly from the primary mafics. Other evidence for the lateness of these changes is in the undistorted pseudomorphs of bastite after orthopyroxene (Pl. 1,

PLATE 6. FELSIC AND BOJITIC SCHLIEREN

In Gabbro. Hillsdale Quarry, Baltimore. Handle and blade = 4″.

FIGURE 1.—HIGHLY CONCORDANT SCHLIEREN

Note fault passing through felsic and mafic material.

FIGURE 2.—DISCORDANT SCHLIEREN

Note transgression of foliation.

fig. 3)—the ultramagnesian rocks were certainly emplaced before these changes began.

The change to chlorite, rather than talc, in some amphiboles is attributed to an addition of alumina by the gabbro during reheating.

Two sets of conditions then obtained—an early relatively dry and a later wetter one (Fig. 16). The presence of plagioclase contributing alumina produced more hornblende and less actinolite-anthophyllite in the early stage (fall of temperature) and more chlorite and less talc in the later (reheating). The higher alumina of clinopyroxenes also allowed a direct transition into hornblende.

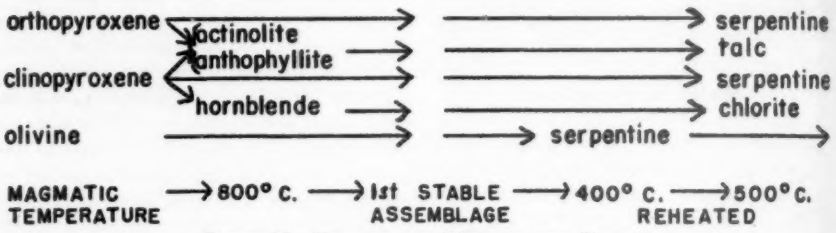

FIGURE 16.—DEUTERIC AND HYDROTHERMAL EFFECTS

Figure 16 fits the ultramafic margin completely; within the gabbro, ultramafics apparently did not get beyond the 400°C. stage. In them serpentine is well developed in the olivines and to a very low degree in the pyroxenes. Talc and chlorite are invariably absent. The suggestion is therefore inherent that they too were reheated after initial crystallization, perhaps by late igneous and hydrothermal solutions that have left so much other evidence as to their presence, as chlorite, albitite, and quartz veins; zeolites; and granite and quartz diorite intrusions.

Contact Effects

Quartz diorite.—Approaching the quartz diorite contact, the gabbro is cut by small veins of that rock indicating that it is intrusive into the gabbro. The actual contact is an area in which the two rocks alternate in bands. In the tunnel, with the contact about perpendicular to the tunnel azimuth, the contact area extended nearly 1000′ with the bands averaging about 50′. The contact appeared conformable.

Contact effects in the gabbro consist of uralization of pyroxene and saussuritization of plagioclase and some chloritization in the ultramafic amphibolites.

Granite.—Contacts with Ellicot City and Port Deposit granites are sharp, with no alternating bands, largely conformable and with little contact effects. Occasional aplitic and alaskite dikes, peripherally arranged about the granite, cut the gabbro, revealing the later age of the granite. Jointed xenoliths of gabbro in granite offer more direct evidence (Pl. 4, fig. 3).

Because of the more intimate relationship between the quartz-diorite and the gabbro than between the granite and gabbro, quartz diorites probably preceded granite. (For other evidence *see* Vol. XIII, Md. Geol. Survey.)

Wissahickon schist.—No systematic change was found in the Ab:An ratio of the plagioclase in schist in the aqueduct, from the Patapsco River to the gabbro contact. At the contact, Ab_{69} occurs; half a mile into the schist, Ab_{70}; 1 mile Ab_{83}; 3 miles, Ab_{82}; 4 miles, Ab_{73}; and at the Portal shaft (6 miles), Ab_{75}. Thus, the variation within the schist, which is entirely in the oligoclase facies, is greater than any effects caused by the gabbro.

Garnet is, however, definitely developed as a result of the gabbro intrusion. In the interior of the schist, garnet is only occasionally seen. Towards the gabbro contact, its frequency is markedly increased, and its crystals more idiomorphic. However, the garnet reveals the same optical and specific gravity data in all samples. Over 5 miles from the contact, N = 1.786 and s.g. 4.02 is found; 3 miles, N = 1.798, s.g. 4.15; 1 mile, N = 1.798, s.g. 4.04; half a mile, N = 1.794, s.g. 3.94; at the contact, N = 1.795, s.g. 4.03.

A garnet is also developed within the gabbro.

Mafic minerals do not show any systematic disposition about the contact. Thus, amphiboles, micas, and chlorites are developed indiscriminately in different areas; oftentimes interlaminated in schists of varying compositions.

Zoning in the schist thus has not been developed as a result of the intrusion; a garnet is a contact effect, however.

Summary and Conclusions

Mapping of the Baltimore gabbro complex, especially in the Baltimore aqueduct has confirmed a general funnel-shape of the intrusion, with the southeast side dipping less steeply than the north and west. Various structures within the gabbro are essentially parallel, invalidating Cohen's differentiation of certain primary and secondary structures. Parallelism of foliation and schistosity is interpreted as indicative of crystallization under tectonic conditions.

Different petrographic groups are distinguished. Mineralogic evidence shows that ultramafic types were the result of strong fractionation within the gabbro magma, and normal crystallization processes including fractionation of wet fractions were responsible for the gabbros and bojites. The brown hornblende of previous authors, allegedly only primary, is shown to have also developed by uralitization.

No metamorphic zoning of any kind was observed within the gabbro. Two plagioclases are present, An_{78-88} and An_{64-70}, but they are primary. All later changes in the mafic minerals, uralitization, serpentinization, chloritization, and steatization, are explained by an initial fall of magmatic temperatures coupled with a later rise.

The sequence of intrusion is thought to have been:

1. Ultramafics emplaced along present borders of the complex.
2. Auto-intrusion of gabbroic facies, with included fractionated bands of ultramafics and wet fractions of bojites. Since these bands are sometimes gradational and never show contact effects, a synchronous intrusion is indicated. Xenoliths, steatitization, serpentinization, and chloritization offer evidence for the earlier age of the border ultramafics.
3. Further differentiation and fractionation producing the quartz-bearing types.

Mineralogical and chemical lines of descent further explain the fact that the crystallization of the Baltimore gabbro complex was entirely in line with established petrologic theory, i.e., from ultramagnesian through gabbroic into quartz-bearing rocks by an initial iron and later alkali enrichment of the magma.

Since the space for the intrusion had already been determined and the gabbro was intruded under tectonic conditions followed later by the more acidic intrusives, an early Appalachian age seems logical. The metamorphism of the Glenarm series would be Taconic under this interpretation.

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Manuscript Received by the Secretary of the Society, August 11, 1950.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 1017-1078, 9 FIGS. SEPTEMBER 1951

PRECAMBRIAN STRATIGRAPHY OF MINNESOTA

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ABSTRACT

This paper summarizes the results of the Geological and Natural History Survey of Minnesota as to the sequence of Precambrian rocks in and near Minnesota.

Eleven formations are noted, and exposures are described to show a well-established sequence. The division planes between groups are also established by four well-studied and described unconformities, two large and two smaller. Other Precambrian formations in the State are more isolated and only tentatively correlated with the main sequence. There have been suggestions also of other unconformities, but the evidences for these are considered doubtful.

The principles and methods of correlation in the Precambrian are briefly noted and applied. Most valuable are the ages determined by radioactivity, the characters of the zircons in large intrusives, the petrographic peculiarities of the iron-formations, and the sequence of beds.

The sequence finally indicated suggests satisfactory correlations of most formations in the State, but only tentative correlations with other series as remote as those north of Lake Huron, the type locality of "Huronian". Most other well-studied series are not so complete as that in and near Minnesota, and more work is needed to make correlation satisfactory.

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Foreword

Several geologists of the Minnesota Geological Survey have had many seasons of field and laboratory study of Precambrian formations. Problems remain, but enough has been done upon an exceptionally complete sequence of rocks to justify publication for use in adjoining and even in remote areas.

The Monographs of the United States Geological Survey on the iron districts of the Lake Superior Region (summarized by Van Hise and Leith, 1911) were written in the early stages of their development; and later Leith, Lund and Leith (1935) summarized important changes, but did not go far into detail as to Minnesota localities in which relations of formations are well exposed.

Several papers suggest correlations with Minnesota formations, and three (Royce, 1936, p. 97–98; 1942; Seaman, 1943) have either ignored or misinterpreted some geologic relations in Minnesota that deserve attention. The chief purpose of this paper is to publish the facts of sequence and to describe exposures fully. A little more than half the State of Minnesota has Precambrian bedrock, exposed or directly under glacial drift (Fig. 1). Most of the localities mentioned are shown on the Geologic Map of Minnesota (1932) still available at the Minnesota Geological Survey.

Thiel is responsible for the writing on the Upper Keweenawan; Schwartz on the Ely greenstone, on the Lower Keweenawan, on the Keweenaw Point volcanics, and on the Thomson formation; Gruner on the Knife Lake group and the Mesabi range; Grout on the Coutchiching, the Soudan, the granites, the Cuyuna rocks, the Sioux formation, and the Middle Keweenawan intrusives. Each writer, however, has worked over the whole report. The members of the staff testify to the helpfulness of Dr. William H. Emmons, the Director from 1910 to 1945, and are grateful for the opportunity to become familiar with this instructive area.

Figure 1.—Counties of Minnesota

Introduction and Relations to Adjoining Areas

The relations of the Precambrian rocks to Cambrian and later sediments are mentioned in the section on the Keweenawan; those of Minnesota Precambrian to the Precambrian of other districts and regions are in a final section.

Most of the group and formation names are based on local exposures described years ago. The exposures in some near-by districts may justify correlation, but for more distant districts and, in the absence of good exposures for adjacent districts, correlation is much less certain.

The sequence here shown for Minnesota rocks differs in certain features from that on the

south shore of Lake Superior and in Canada. We agree that certain correlations can be carried from Minnesota to Michigan, but are not clear that the term Huronian is safely applied.

of more than two eras. Schuchert and Dunbar (1933) suggest that the whole Precambrian be called a Cryptozoic Eon, presumably composed of several eras, analogous to a Phanerozoic

TABLE 1.—SUGGESTED NAMES OF MAJOR DIVISIONS OF THE PRECAMBRIAN

	Adams (1909)	Van Hise and Leith; USGS (1911)	Grabau (1921)	Schuchert and Dunbar (1933)		Wilson (1939) and Canadian Geol. Survey (1946)		Stose, GSA Map (1946–1947)	Pettijohn (1937)	
Precambrian	Neo Proterozoic	Algonkian	Proterozoic	Proterozoic	Late Cryptozoic	Proterozoic	Late	Later	Late	IV
	Meso Proterozoic		Eozoic	Archeozoic	Middle Cryptozoic		Early	Medial	Early	III
	Eo Proterozoic	Archean	Archeozoic		Early Cryptozoic	Archaean		Earlier		II
										I

TABLE 2.—GRANITE BATHOLITH INVASIONS NEAR LAKE SUPERIOR AS INDICATED BY STUDY OF ZIRCONS

South Shore	*North Shore*
Granite (Keweenawan)	Granite (Keweenawan)
Keweenawan surface rocks	Keweenawan surface rocks
Granite (malacon-bearing)	
Huronian? group	Animikie group
Granite (malacon-bearing)	(Some recent findings)
	Granite (hyacinth-bearing)
	Knife Lake group
Granite (hyacinth-bearing)	Granite (hyacinth-bearing)
Keewatin group	Keewatin group

A complete time sequence for the Canadian Shield, or even for the Lake Superior region, may include groups and formations not yet known in Minnesota. Leith, Lund, and Leith (1935) say "special studies of this complex have from time to time resulted in the separation and better definition of geologic units, and this breaking up is likely to go further."

ATTEMPTS TO BUILD A GEOLOGIC COLUMN

Table 1 indicates the trends of opinion in the last 40 years. Important unconformities appear at two or more horizons. By past custom one has been selected as a "great" unconformity, giving a dual nature to the Precambrian; but the trend is clearly toward the recognition

Eon composed of Paleozoic, Mesozoic, and Cenozoic Eras. The Cryptozoic is probably at least three times as long as the Phanerozoic.

Another sort of study is based on such different features that we have not tried to incorporate it into Table 1. The work on heavy accessory minerals, especially zircons, led to tentative conclusions as to the number of Precambrian batholithic invasions, which remarkably support the larger number of revolutions inferred in Table 1.

Table 2 is based on the finding (Tyler *et al.* 1940) that (1) Archean and Algoman granites carry "hyacinth" zircons, (2) Late Huronian granites carry malacon with or without some older hyacinths incorporated, and (3) Keweenawan igeneous rocks carry "normal" zircons

with or without some of the older kinds. The three varieties of zircon were described in detail, and the method was checked in the field by many masses of known age.

The sequence for Wisconsin and Minnesota is shown in Table 2. The correlation of a pre-Huronian malacon-bearing granite of the Wisconsin area with the Algoman of the Minnesota area may well be left open for further study.

The major terms of Table 3 follow those of the recent geologic map of North America. The unconformities known in Minnesota are described in later pages and summarized in a note on their relative importance. They show no points favoring the simplicity of a dual division. The unconformity at the base of the commercially important Animikie iron-bearing group, however, is traced widely in Minnesota, Ontario, Wisconsin, and Michigan. Others are not so widely recognized.

The plain fact is that Precambrian time is long enough for half a dozen such eras as are recognized in later geologic history (Holmes, 1931, p. 440–441). The path of progress is to recognize the divisions and sequence wherever careful work is done. Ultimately it may be possible to correlate more of the groups through near-by districts if not through great regions. A study of districts near Minnesota confirms the suggestion that the total sequence includes more real eras than are recognized here. Pettijohn (1937) finds, a few miles north of the International Boundary, good evidence of two granite intrusives and unconformities below the Algoman interval, whereas Minnesota has evidence of only one.

All attempts at a dual classification make one era straddle a great unconformity. We prefer a three-fold (or four-fold) division. The three-fold division, used by the recent geologic maps of Canada and of North America, puts in the "Later Precambrian" both Animikie and Keweenawan; while these are separated at several places by unconformity, it may not be as great a break as the others.

If names are wanted for three eras, there are the suggestions of Table 1; Seaman (1943) prefers Azoic for the oldest, and Dawson (1868) proposed "Eozoic" for the age of *Eozoon Canadensis*. These, however, imply a more widespread correlation of beds and recognition of major unconformities than has yet been established. The less definite names used here are preferred as not having the sound of well-established divisions.

Names for the formations and members are mostly local as is generally approved. Van Hise (1892, p. 474) said "the tendency to return to the use of local names is plainly a reversion to scientific methods."

Note should be made that we use "eras" where we have at least an approximate idea of widely separated age limits, as will be noted in description of the Algoman and pre-Knife Lake granites. If the Cambrian began about 500 million years ago, the Algoman granites are dated about 750 million years ago, and the old granites near Lake of the Woods 2300 million years ago, we have three eras definitely bounded except for the beginning of the oldest one. The time between these two granites is long and may need subdivision. Possibly a word should be added as to the time limits of the proposed eras. The real determinations used here are the ages of certain minerals formed late in periods of batholithic invasions, and these are only roughly the same as the unconformities used as division lines in Table 3.

In naming the rocks in sequence in each of the three eras we refer to *groups* and *formations* rather than *periods* and *epochs*, or *systems* and *series*, because the time limits are not yet as well determined as the limits in post-Keweenawan times, in which there are fossil evidences.

No Coutchiching Recognized in Minnesota

Lawson (1887) named and described the supposed relations of the Coutchiching on the basis of exposures largely on the shores of Rainy Lake. A committee of geologists, representing Canadian and U. S. Geological Surveys, visited the type locality in 1905, and Hays *et al.* (1905) reported that the schists called Coutchiching are stratigraphically higher than those mapped as Keewatin. Lawson restudied the area and again reported Coutchiching (1913). After a study of several critical areas Grout (1925a) reported that structural features such as cross bedding, gradational bedding, and drag folds indicated that all the Minnesota localities gave evidence that the sediments were younger than the Keewatin greenstones. Even if there are, on the Canadian

TABLE 3.—PRECAMBRIAN COLUMN IN MINNESOTA CHRONOLOGIC AND STRATIGRAPHIC SEQUENCE

	Eras and Rocks	Groups	Formations and Members	Other Names Used in Minnesota
Phanerozoic Eon	Cenozoic Mesozoic Paleozoic	(For detail *see* Minn. Geol. Survey Bull. 29)		
		unconformity, great on Wisconsin boundary and east		
Precambrian (or Cryptozoic Eon)	Later	Keweenawan Group — Upper: Sandstones and other sediments	Hinckley Fond du lac	Red Clastics*
		Keweenawan Group — Middle: Intrusives, acidic and basic	Scattered granites* Duluth gabbro Beaver Bay Complex, and Logan intrusives	
		Keweenawan Group — Middle: Flows, tuffs and sediments	Keweenaw Point Volcanics	
		Keweenawan Group — Lower: Conglomerate and sandstones	Puckwunge formation	
		unconformity, may be great on the Cuyuna range		
		? Sioux formation* ?		
		Animikie Group	Virginia slate (minor iron formation?) = Rove = Biwabik iron formation series = Gunflint = Pokegama quartzite	Upper Cuyuna slate* Deerwood*
		unconformity, great on the Mesabi and Gunflint ranges		
	Medial	Algoman batholithic intrusives, orogeny and erosion	Central and southwestern Minnesota granites*	
		Knife Lake Group (about 18 members)	slate = Thomson* graywacke iron-bearing beds conglomerate tuffs, lavas and intrusives	(Carlton)* Emily* Agawa* Ogishke*
		unconformity, great west of Saganaga Lake		
	Earlier	Pre-Knife Lake batholithic intrusives, orogeny and erosion		
		Keewatin volcanics, and Soudan iron formation member. (No Coutchiching recognized in Minnesota)		Ely greenstone

* Means uncertain as to place.

side of the boundary, some sediments below greenstone, it is impossible to state their relation to older rocks, and their top varies in relation to rocks of several different ages. The formation can hardly be defined.

The descriptions were questioned and discussed in several papers largely on the basis of exposures in Canada (Bruce, 1927).

After a detailed study east of Rainy Lake, Merritt (1934) found that most of the sediments are younger than the Keewatin. He explained by a great fault some exposures that had been supposed to lie below greenstone and is convinced that the sediments on the Minnesota side of Rainy Lake are younger than Keewatin. Pettijohn (1937) made a broader study. North of Rainy Lake he finds greenstones, probably of different ages, some above sediments. Even the oldest greenstone may, in some Canadian districts, have underlying sediments—and these may be called Coutchiching, even though the similar schist "of the type locality is much later in age." These recent papers give references to and summaries of the rather extensive literature.

The two localities in Minnesota at which Coutchiching rocks in the original sense—older than Keewatin—might be suspected are at Rainy Lake and at Burntside Lake. Grout (1925a, Fig. 2) sketched Burntside and the major area of Rainy Lake. These probably need no revision.

The structures were reviewed by the international committee (Hays *et al.*, 1905), by Merritt (1934), and by Pettijohn (1937) with general agreement. Tanton has not, at the last reports, accepted the conclusions. Maps are here offered of critical spots where evidence is good.

An extensive greenstone belt crossing Grassy Island (Fig. 3) is generally agreed upon as Keewatin. Close to the greenstone are green slates and mica schists. While some of the sediments give structural evidences of the attitude of the tops of the beds, most of the green slates do not. It is here that Tanton is most in disagreement with other students.

North of Grassy Island is Jackfish Island with green sediments near the greenstone and biotite schists. The biotite schists north of the green rocks show gradational bedding with tops of beds toward the north (Tanton claims to have seen opposing structures). Southeast of Grassy Island, on the mainland between Frank Bay and Neil Point, is conglomerate and arkose sandstone whose cross-bedding shows that all across Neil Point the tops of the beds are to the south. South of the Point is a narrow belt of green sediment much like the matrix of the conglomerates; farther south is a mica schist. Clearly these structures, one north and one south of the greenstone belt, if confirmed, make the mica schist younger than the Keewatin along the Minnesota shore. Most students agree to this, and anyone can find them as indicated in Figure 3.

The anticline Lawson (1913, p. 10) thought he found north of Bear Passage may well be attributed to a doming of the sediments by the granite stock south of Bear Passage. The stock is about 2 miles across though partly concealed under the lake. Nevertheless the strikes of much of the surrounding sedimentary structure suggest the structural pattern of a dome. South of the granite is Morton Island, whose rocky shores are washed clean by the waves. The sedimentary structures are not so much obscured by metamorphism as are those Lawson described along the railroad. It seems worthy of detailed study.

Figure 4 is a map and an inferred structure section of Morton Island. Grout and Tanton have agreed as to the exposures and attitude of the tops of the beds, but not to the interpretation. Grout believes that the sediments lie above the greenstone, which is better exposed in the larger island to the south. This is based on the structure section and the green color of the most southerly sediments, whereas the green beds are thinner and less abundant toward the northwest side of Morton Island. In general, the structure shows close folding, but the beds tend to go deeper toward the northwest, toward the granite of the stock. This opposes the idea that the granite lifted the sediments in any simple dome. Probably the granite rose into closely folded beds, more or less disregarding the structure. Even an overturned bed can be domed by a rising granite.

Earlier Precambrian Era And Rocks

General Statement

The oldest rocks known in Minnesota are variously called Archean, Archeozoic and

Azoic. Since these terms are widely used it may seem ultraconservative to maintain that we do not know that the old rocks are equivalent to those called Archeozoic elsewhere. The next system in Minnesota has been variously assigned to the Archeozoic and to a later system. Thus if we used the term we would have to define its upper limit arbitrarily or admit that it was at an uncertain horizon.

The Earlier Precambrian System as here used includes the Keewatin Period, and the time of the following period of invasion, orogeny, and erosion. The rocks include (1) Keewatin flows and associated Soudan iron formation and minor sediments, and (2) the pre-Knife Lake granites, usually designated Laurentian.

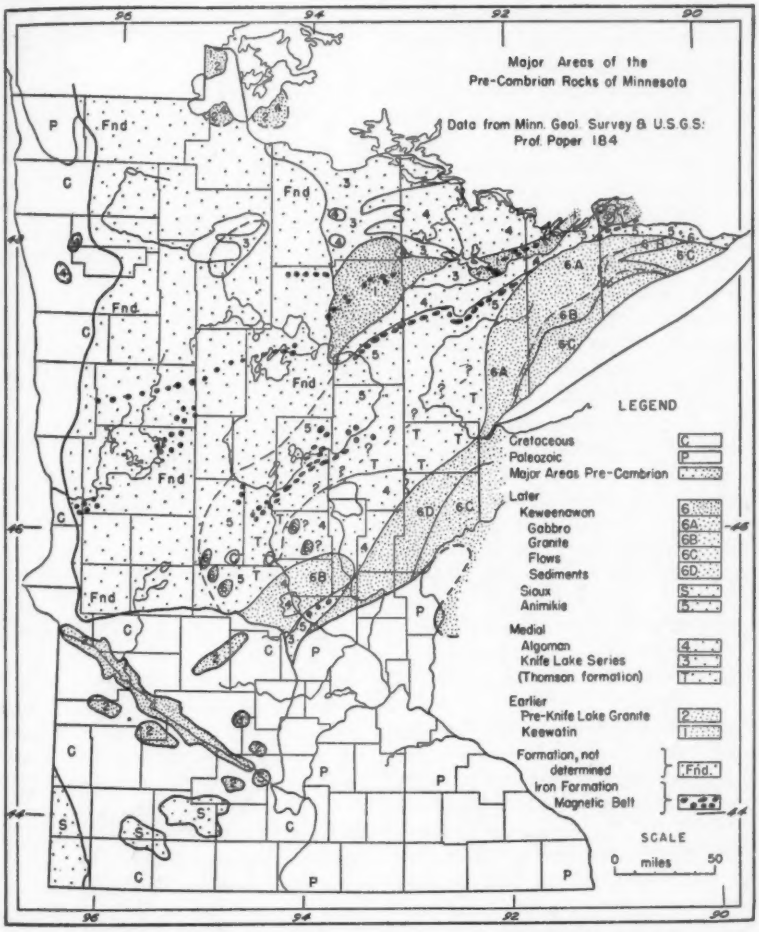

FIGURE 2.—MAJOR AREAS OF THE PRECAMBRIAN ROCKS OF MINNESOTA

Keewatin Group

Definition.—Lawson (1885) named the Keewatin from exposures on Lake of the Woods, most of which are in the Keewatin district of Canada. There is no doubt, however, that the series is equivalent to outcrops south of the boundary (*See* Leith, Lund, and Leith, 1935, Pl. I). It runs east on the Canadian side nearly to Port Arthur and almost connects with the

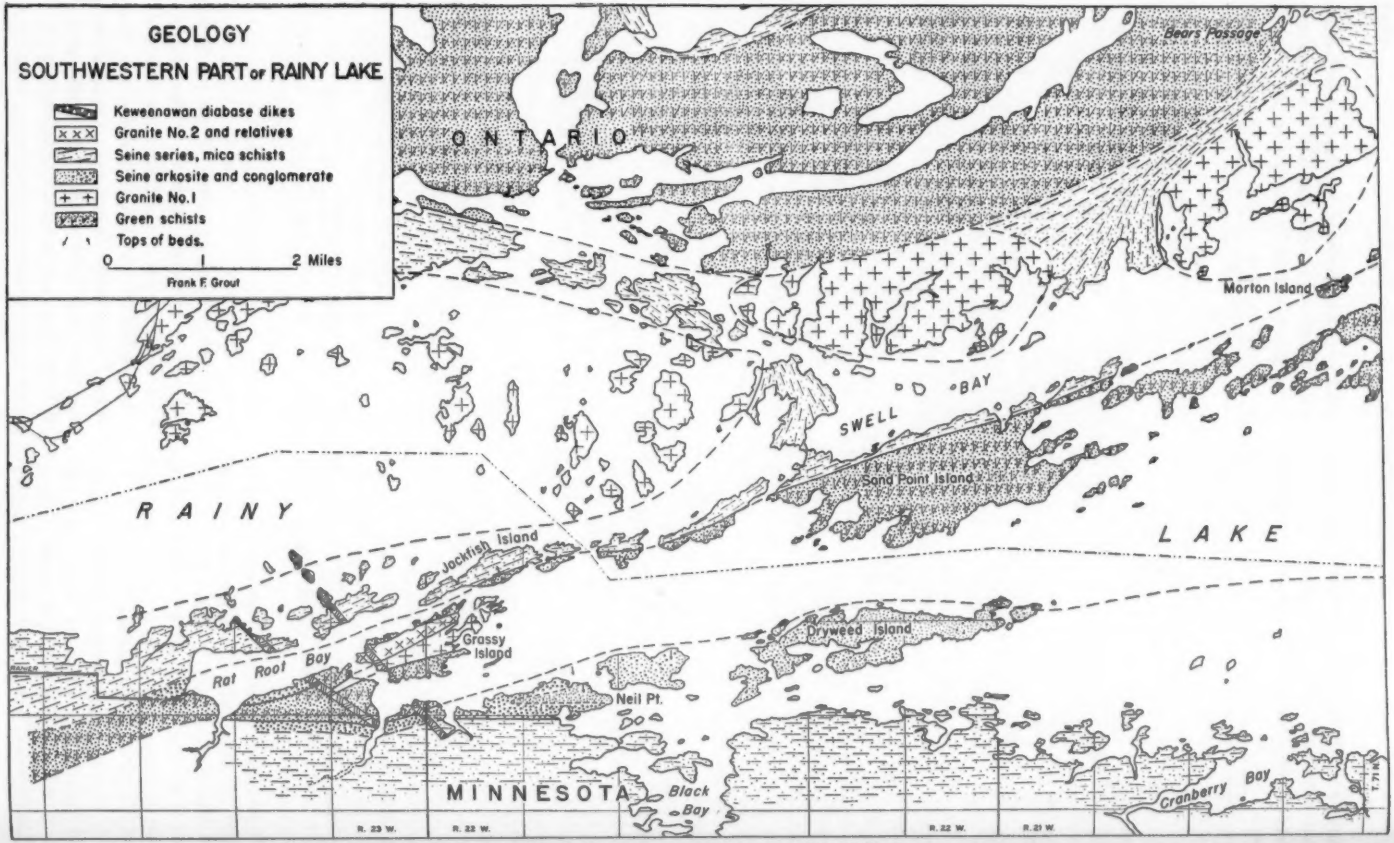

FIGURE 3.—GEOLOGY OF A PART OF RAINY LAKE

Minnesota belt from Saganaga Lake to Cass County. West of that, drift cover is thick.

The joint committee (Hays *et al.*, 1905) found the original Keewatin a fairly well de-

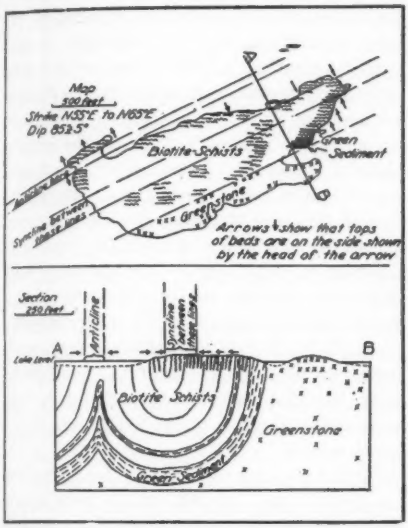

FIGURE 4.—MAP AND SECTION OF MORTON ISLAND, RAINY LAKE, SOUTH OF BEARS PASSAGE

Where Coutchiching sediments have been thought by some to lie below greenstone.

fined unit and approved the name to cover greenstone, green schist, and the iron-formation of the basement complex, apparently considering it below the Knife Lake (Seine) group. It is divided in Minnesota into the Ely greenstone and Soudan iron-bearing formations.

Leith, Lund, and Leith (1935, p. 9) describe the Keewatin as basaltic flows (with thin slate and iron-formation) "unconformably beneath all unquestioned Huronian sediments as well as the Knife Lake series and its correlatives."

Ely greenstone.—Van Hise and Clements (1901) named the Ely greenstone, and Clements (1903) described the formation in detail. The greenstone and associated iron formation had been previously termed the Kawishiwin Series by Winchell (1899b, p. 270) who supposed that some metamorphosed facies of the Knife Lake group were the upper part of the Keewatin. The practice of the United States Geological Survey to separate the greenstone and an iron-formation member has been followed since 1901. Clements (1903, p. 303) shows them unconformably below the Knife Lake group. The type exposures occur in the city of Ely, and mine shafts have penetrated a considerable thickness. The greenstone is fairly resistant to erosion and forms prominent dome-shaped hill which rise 100 to 200 feet above adjacent areas of metasediments and granite.

LITHOLOGY: The formation is not uniform, and the various areas are not necessarily of the same age. Some rocks are effusive, others intrusive, and still others fragmental, but nearly all are much altered and green because of abundant secondary chlorite.

Original differences and more especially subsequent alteration have developed a number of petrographic, as well as structural, varieties, chiefly ellipsoidal, spherulitic, schistose, tuffaceous, siliceous and diabasic.

Ellipsoids are common and are the safest criterion for recognition of the greenstone. They range from less than an inch to 3 or 4 feet in diameter. Locally they are flattened. The ellipsoids and matrix have been described as essentially similar mineralogically, but there are exceptions. The detailed petrography is described by Schwartz (1924).

The best exposure of tuffaceous materials noted in the field is on the north shore of Knife Lake at the narrows in Section 24, T. 65 N., R. 7 W.

Strongly schistose greenstone is common only near the later granite intrusives. The entire formation has been referred to as a schist, but normally the schistosity is only imperfectly developed.

MINERALS OF THE GREENSTONE: The most abundant minerals are plagioclase, chlorite, hornblende, epidote, calcite, magnetite; of lesser importance are leucoxene, quartz, actinolite, augite, sericite kaolin, limonite, pyrite; rare minerals are zoisite, biotite, bronzite, talc, apatite, sphene, perovskite, and ilmenite.

The Ely greenstone is intruded by the Saganaga, Vermilion, and Giants Range granite batholiths. At and near the contacts metamorphism is noteworthy, commonly a gradation from normal greenstone to a hornblende schist. Good contacts may be seen along a road running south about a mile east of Ely;

FIGURE 3.—GEOLOGY OF A PART OF RAINY LAKE

on the portage in Sec. 24, T. 63 N., R. 10 W.; and on the portage from Jasper (Frog Rock) Lake to Alpine or West Gull Lake (Sec. 17, T. 65 N., R. 5 W.).

The Duluth gabbro intrudes the greenstone along the southeastern border of the Vermilion range, forming a hornfels of sugary texture. The abundant minerals are recrystallized plagioclase, augite, hornblende, and magnetite. The best exsposure of the gabbro-greenstone contact is at Disappointment Mountain (Sec. 35, T. 64 N., R. 8 W.). The greenstone is also exposed near the contact at Gabimichigami Lake (Sec. 1, T. 64 N., R. 6 W.).

DISTRIBUTION: The principal exposures of the Ely greenstone follow the Vermilion iron-bearing district from Tower to Gunflint Lake. The largest, including the type locality, extends 40 miles from Vermilion Lake to Moose Lake (T. 64 N., R. 9 W.). Another extends northeast into Hunter's Island, Canada. East of Moose and Snowbank lakes on the Vermilion range, the formation occurs as isolated areas in the Knife Lake group or in contact with the gabbro and granite. Gruner (1941) has shown that a few small areas mapped as Ely by Clements (1903) are probably later intrusives and extrusives involved in the complex Knife Lake group.

West of Tower, outcrops are not numerous, and the boundaries in western St. Louis County have been estimated by Grout (1926, Pl. 1). In northwest St. Louis County, Grassy Island of Rainy Lake exposes Keewatin green schist forming part of a belt extending into Koochiching County. In Itasca County west of the main belt of St. Louis County, the greenstone is known in drill holes, as at Northome (T. 151 N., R. 28 W.) and in scattered outcrops, as at Kelliher (T. 152 N., R. 30 W.).

Outlying areas of greenstone, correlated with the Ely, occur at places along the north side of the Mesabi Range. The Keewatin of the type locality, Lake of the Woods, is well correlated with Ely greenstone by belts on the Canadian side of the boundary. Other outlying areas of greenish intrusive diorites and related rocks are tentatively classified as Keewatin greenstone.

STRUCTURE: The fact that the Ely greenstone is composed mainly of altered basalt flows makes it difficult to recognize structure. The careful mapping by Clements (1903) and the very detailed work on the Knife Lake slates by Gruner (1941) show that the structure of younger formations is very complex, and the greenstone, as shown by its relations to these formations, has been involved in the same deformation. As pointed out by Clements the belts of greenstone in the Vermilion district are generally anticlines with the synclines occupied by the younger sediments. The steeply plunging structures of the sediments indicate cross folding which is responsible for the separation of the belts of greenstone. The maps by Clements (1903) show the general structural relations for the district, and Gruner's map shows the detailed structure from Snowbank Lake to Seagull Lake (from T. 64 N., R. 8 W. to T. 65 N., R. 5 W.). Faulting is much more important than was realized before Gruner's work. Fault contacts between the greenstone and several members of the Knife Lake group are numerous.

CONTACTS AND RELATIONS: The Ely greenstone has usually been considered the base of the earlier division of the Precambrian. There is the possibility of a formation below the Ely in adjacent areas in Canada, but in Minnesota the base of the Ely has not been observed, and no sedimentary beds appear to be older than the Ely.

At many places the lower contact of the greenstone is with intrusive granites and gabbro. The Soudan iron-formation lies near the top and partly within the greenstone. Details are given in the discussion of the Soudan.

The greenstone is overlain at places by the Ogishke conglomerate with abundant greenstone pebbles and even boulders. Much conglomerate earlier classified as Ogishke has been shown by Gruner (1941) to be in the Knife Lake group rather than basal conglomerate at the unconformity above the Ely greenstone. Gruner describes the conglomerates at Disappointment Mountain (Secs. 26, 27, 34, 35, T. 64 N., R. 8 W.) and south of Moose Lake (T. 64 N., R. 9 W.) as containing a predominance of greenstone fragments. A belt of Ely greenstone extends from Jasper Lake (Sec. 1, T. 63 N., R. 10 W.), and the Lake-Cook county line, eastward about 10 miles to Section 27,

T. 65 N., R. 4 W. Along this belt greenstone is overlain by conglomerates high in the Knife Lake group which gradually pinch out eastward. The Animikie (Gunflint or Biwabik) iron-formation, which lies unconformably on the Knife Lake group along the west part of this belt, lies directly on the Ely greenstone east of Section 27, T. 65 N., R. 4 W. The Ely greenstone is intruded by the pre-Knife Lake (Saganaga) granite, and, as noted above, the metamorphism is well shown at the south end of Alpine Lake. The granite is in contact with the greenstone for 8 miles to the east.

The Algoman granites intrude the Ely greenstone in a belt north of and eastward from Vermilion Lake past Burntside Lake to Basswood and Wind lakes. The Duluth gabbro intrudes the greenstone along the south border at the Kawishiwi River south of Twin Lakes, at Disappointment Mountain, and at Gabimichigami Lake. Small dikes of Keweenawan basalt are related. Intrusive contacts thus form an important percentage of the total contacts.

AGE: Assuming that the Ely greenstone is equivalent to the greenstone of the Keewatin at the type locality, Lake of the Woods, as it seems to be, the age of the greenstone is very great—greater than that of some pegmatites that intrude it (Ahrens, 1947; 1949, p. 258).

THICKNESS: There seems to be no record of any estimates of the thickness of the Ely greenstone, and its total thickness cannot be estimated unless some formation below it is clearly recognized. The uniformity and lack of bedding in greenstone, and the complex structure indicated by adjacent formations increase the difficulties. Along the north side of the Saganaga granite, the wall rocks of the batholith are greenstone schists, for about 12 miles, with large masses of hornblende schist in the gneiss below. Probably (Grout, 1929b) this long exposure was once a contact in nearly vertical position so that the greenstone walls were nearly 12 miles deep, but no doubt they were folded along the sides of the granite.

Considering the width of the main belts of greenstone and the general structure it seems safe to conclude that the greenstone must consist of many thousand feet of lava flows.

Soudan (iron-bearing) formation.—The Soudan formation was named by Van Hise and Clements (1901) from exposures at Soudan Hill, near the company town of Soudan, Minnesota, and was described in detail by Clements (1903) and briefly by Van Hise and Leith (1911).

Westward and southwestward from Lake Vermilion, beyond the limits of the Vermilion map, Keewatin greenstones, Algoman granite, and Animikie formations have been traced for a considerable distance. An iron-bearing formation, correlated with the Soudan, forms a considerable belt extending from Ts. 60 and 61 N., R. 22 W., southwestward to T. 58 N., R. 27 W. It is sparsely exposed and is known principally by its disturbance of the magnetic field. A small amount of exploration has been done on this belt, but most of the cores seem to be lean and unpromising. On the geologic map of Minnesota the belts are shown as rows of brown dots.

The remarkably complex folding partly explains the distribution of the Soudan formation with reference to the Ely greenstone. Upon the major folds are superposed secondary and tertiary folds, so that numerous patches of the jasper occur in the greenstone. Moreover, because of the cross folding, these patches may be very narrow at one place, widen out within a very short distance to make a thick formation, and narrow again. When the extraordinary complexity of this folding is understood it is only necessary to premise an erosion extending to different depths in the Soudan formation before the Knife Lake group was deposited in order to see how in the greenstone there may be patches of jasper ranging from a few feet in width and length to the dimensions of a great continuous formation about Tower and Ely.

LITHOLOGY: The iron-bearing Soudan formation comprises two classes of rocks. To all the varieties of the first the miners apply the name "jasper." Although only a portion of it falls strictly under this designation, this is the dominant rock. Clements calls the second variety mainly slaty but in some places conglomeratic.

The chief varieties of the "jasper" are (1) the cherty, (2) the black-banded, (3) the red-banded, and (4) a banded variety with white quartz. With these are subordinte masses of (5) the carbonate variety and (6) the ore bodies.

ORIGIN AND RELATIONS: The Soudan formation seems to be above the main mass of the Ely greenstone. In certain places it is in plunging folds, as is the greenstone forming the walls and bottom, at Ely and Soudan.

Some of the jasper belts of the Vermilion district are clearly interbedded with successive basalt extrusives. Such beds, but a few feet thick, may be traced for hundreds of yards with uniform widths and even contacts. Close examination of the adjacent igneous rocks reveals that the sedimentary bands parallel the tops and bottoms of separate flows, as marked by amygdaloidal and chilled surface textures, without intervening fragmental sediments. This is well illustrated in secs. 13 and 14, T. 62 N., R. 13 W., Minnesota.

Because the greater masses of the Ely greenstone were deposited before the larger masses of the Soudan formation, it is believed that the great volcanic period of the Ely greenstone had almost ceased before Soudan time. However, the extremely intricate relations and apparent interstratification of the minor masses of the Soudan formation with the Ely greenstone, and the fact that both the Ely and Soudan formations locally contain interstratified fragmental material, lead to the belief that volcanic activity had not entirely died out.

The relation appears to differ slightly at different places—for example, in the well-explored productive districts at Soudan and at Ely. The geologists of the Oliver Iron Mining Company report that in the "Ely trough" the iron-formation lies near the top of the greenstone, but at least some thin beds of iron-formation occur at more than one horizon in the greenstone.

What were the physical conditions which permitted the deposition of the nonmechanical Soudan formation upon the Ely greenstone with so insignificant an amount of intervening mechanical sediment and erosion surfaces? It is impossible to believe that the sea could advance over rough lava flows without the production somewhere of mechanical sediments of considerable thickness. Van Hise and Leith (1911, p. 600) suggest that the eruptions of the Ely greenstone were submarine, as evidenced by ellipsoidal textures. After the period of volcanism had ceased nonmechanical sediments of the iron-bearing formation might at once be deposited, provided the conditions were proper. The geologists of the Oliver Mining Company think the water may have been in local ponds rather than in the sea.

Several exposures on the bare hills north of Tower show relations of Soudan formation to the Knife Lake conglomerates. Pebbles of jasper, slate, and chert in the conglomerate have been derived from the iron-formation which immediately underlies the conglomerate.

With the clear evidence that the Saganaga granite is younger than greenstone, it is at least very probable that some of that pre-Knife Lake granite intrudes the Soudan formation also, but no exposure has been reported to prove it.

The granite dikes near the Algoman Giants Range granite in secs. 3, 4, 7, and 18, T. 62 N., R. 12 W., and sec. 28, T. 62 N., R. 13 W., are generally assumed to be apophyses from that intrusive because they are most abundant near the granite, and few occur more than 2 miles away. Similar dikes cut the iron-formation in sec. 10, T. 63 N., R. 10 W., but these are about 2 miles from the continuous mass of the Giants Range granite. Clements (1903, p. 199, 359) recognized some of these as related to the Giants Range but confused others with the older granite, which he called Archean.

Dikes near the Vermilion granite, the Snowbank stock, and the Kekequabic stock are probably of the same age and may cut the Soudan formation, but their age is not proved. Such dikes cut the Soudan formation north of Mud Creek Bay in S.E. ¼ sec. 1, T. 62 N., R. 15 W., and for 1–2 miles east of Vermilion Lake, south of Mud Creek.

Porphyry dikes which cut the greenstone and Soudan formation, as they cut greenstone near the water tower at Ely, may be related to the Algoman granite but are petrographically like some of the intrusives and flows of the Knife Lake group.

Since the Animikie lies on the greenstone, and since Keweenawan dikes cut the greenstone, these formations are no doubt in similar relation to the Soudan, but outcrops have not been seen to prove it.

AGE: The Soudan formation is in general younger than the Ely greenstone, the oldest

rock of the district, but on the whole so intimately associated with it that the two must be considered as belonging to the same great period—the Earlier division of the Precambrian.[1]

THICKNESS: It is impossible to make any reliable estimate of the thickness of the Soudan formation, but, as inferred from its superficial extent rather than from any definite measurements, it is presumed to reach several hundred feet.

Pre-Knife Lake Intrusives

The term Laurentian is widely used but variously defined. It was originally used by Logan (1854) referring to the exposures of gneiss, granite, and other old rocks north of the St. Lawrence River. The special committee (Hays *et al.*, 1905) approved the term for "granites and gneissoid granites which antedate or protrude through the Keewatin, and which are pre-Huronian;" also for some granites of unknown age. Cooke (1920) urged that the term Laurentian be reserved for those granites that cut *all* rocks below the base of the Huronian, and that other terms be found for those that cut only the Keewatin. The tendency since that time even among Canadian writers is to use the term Laurentian for the oldest granites.

Radioactivity determinations may show seven or more Precambrian granites (Holmes, 1931), but in Minnesota we recognize three groups: the first *post-Keewatin and pre-Knife Lake;* a second (Algoman) post-Knife Lake and pre-Animikie; and a third Keweenawan. The term Laurentian is commonly used for the oldest group (our pre-Knife Lake granites), though some of them may be older than others (Burwash, 1933).

The oldest granite near Minnesota is probably the gneissic one on Northern Light Lake, east of Saganaga Lake on the boundary (near T. 65 N., R. 2 W.). It is largely a migmatite of granite with hornblende schist that seems to be derived from greenstone, and the granitic component is a sodic granite (Grout, 1929b). This oldest granitic intrusive is not recognized in Minnesota, but some gneisses associated with younger granites may be of this age.

The major mass of definitely pre-Knife Lake granite is the Saganaga granite cropping out conspicuously on the shores of Saganaga Lake. Its structure and relations have been mapped in some detail (Grout, 1936). A very striking petrographic feature is the occurrence of quartz in patches up to half an inch across, suggesting phenocrysts (though microscopic study shows that they are aggregates). This is the only granite in the State with much of this peculiarity, which thus serves to identify the mass as a petrographic unit some 15 by 25 miles across. The foliation and lineation shown by quartz and hornblende, by inclusions of greenstone and by schlieren dikes, all confirm the impression that the whole area acted mechanically as a unit. Its deepest portions are petrographically much like the rock at its upper contacts, but the deeper (eastern) exposures suggest a series of fingers, which may have been roots, or feeders, of the main batholith.

The petrographical and structural unity of the mass is emphasized here because Tanton's map (1933) shows Laurentian only at the west edge of the mass where it is close to the contact with Knife Lake sediments—the rest of the area is shown as Algoman. The quartz grains are generally coarse through miles of exposures, and finer (or almost absent) near some borders. Very few are related to fractures or veins. The overlying conglomerate has some boulders with, and some without, such quartz. The writer has shown a number of geologists over the area, and among those who have studied the area since 1891 only Tanton thinks that any of the granite with coarse quartz is younger than the conglomerate of the Knife Lake sediments. The granite is clearly a unit, and older than the Knife Lake group.

The major facies, making up about 80 per cent of the exposures of Saganaga granite, is a hornblendic pink granite with enough oligoclase so that Johannsen (1932, p. 334), basing his division points for plagioclase at An_{10} rather than An_{20}, classes it as granodiorite. Border facies have much less quartz and grade to

[1] Royce (1935; 1945) has suggested that the Vermilion range ores may be of the same age as the Mesabi ores. The easily accessible outcrops here described disprove such an idea.

hornblende syenite, shonkinite, and hornblendite in a small discontinuous zone. The foliation and lineation mentioned are accompanied by a joint system (commonly followed by aplite dikes) at right angles to the lineation. Details of petrography and analyses are given by Grout (1929b).

There are good exposures of the deeper parts or "roots" of the mass on the Canadian side of the boundary. The older gneiss is cut in a few places by dikes of granite as, for example, on a small island, about 5 miles east of the outlet from Northern Light Lake where it discharges into Saganaga Lake.

The intrusive relation of Saganaga granite to greenstone is well shown in the west half of the mass, along both the north and south sides. At Alpine Lake, secs. 7 and 17, T. 65 N., R. 5 W. (Minnesota), many dikes large and small cut the greenstone. Granite dikes cut and brecciate the greenstone on the Canadian side, from the northwest part of Cache Bay at many places to the end of the northeast arm of Saganaga Lake. Fragments of greenstone, as inclusions a few inches to a few feet across, are widely scattered in the granite.

The upper contact of the granite is at its western edge where Knife Lake sediments rest upon eroded granite and greenstone. At the west side of Alpine Lake (Sec. 7, T. 63 N., R. 5 W.) the conglomerate resting on the granite carries not only granite pebbles but some of greenstone and some of the red jaspilite characteristic of the Soudan formation. The contact as mapped is so intricate as to indicate a certain amount of faulting after sedimentation.

The contact and the beds of sediment dip about 70° W. Immediately east of the contact the granitic quartzes show orientation in a pattern consistent with that through the whole mass. Locally along the international boundary the linear features in the granite plunge about 15° E. Some aplites in cross joints dip 75° W. (almost parallel to the bedding of the overlying arkosite, but the textures are not to be confused).

North of these outcrops, on the west side of Cache Bay, the granite contributed coarse boulders to a thick conglomerate, which grades westward into cross-bedded sands (tops to the west), and, in the small inlets that extend farthest west, the arkose sands grade into varved slates and graywackes (tops to the west).

Formations younger than the Knife Lake group also occur in contacts showing they are younger than Saganaga granite. Basal Animikie conglomerate rests on old granite a few rods north of "Railroad Bay" on the north shore of North Lake which is in T. 65 N., R. 3 W. Several trap or diabase dikes, probably of Keweenawan age, cut the granite on the shores and islands of Sea Gull Lake and along the portage from Sea Gull to Saganaga Lake.

Several other granite masses in Minnesota may be earlier than the Knife Lake group, but none is in contact with sediments which prove its age. Several granites at Lake of the Woods are in contact with no rocks but greenstone. Ahrens (1947; 1949, p. 258) by radioactivity methods finds the tentative age of these granites to be 2100–2350 million years.

The many granites of the southwestern part of the State lack exposures of contacts with sediments of known early Precambrian age. Cretaceous sands and clays lie on them unconformably (Grout, 1919, p. 218). The zircon crystals of many of them suggest that they are either Laurentian or Algoman. State maps have generally designated them as "Archean" if they looked old. Lund (1950) has found some that contain malacon and seem to be "pre-Huronian or early Huronian" (Table 2).

Post-granite Unconformity

The unconformity between the pre-Knife Lake granite and the Knife Lake group involved time enough to cut through the Keewatin roof of the Saganaga batholith to an unknown depth into the granite itself. Great boulders of granite were transported as far as 20 miles to be deposited in conglomerates near Ensign Lake. Since the intrusion of the batholith probably involved mountain-making folds in the thick Keewatin greenstones, the total erosion was probably many thousands of feet, perhaps as much as 5 miles (Grout, 1936, Fig. 9).

One series of exposures north of Gunflint Lake, which is in T. 65 N., R. 3 W., presents a problem that may involve events in this interval, but the exposures are on the Canadian side of the lake and not very extensive, so it may be difficult to solve. Between the Animikie, at the shore of Gunflint

Lake, and the granite ridge about a mile north, are several rocks older and more metamorphosed than the Animikie. Some are greenstones which appear to be Keewatin, but others are black slates of uncertain relation to the greenstones—they resemble the Knife Lake slates. The problem is that both the greenstone and slate are intruded by acidic porphyry dikes such as might be apophyses from the Saganaga granite; in the district no slate is known to be older than the Saganaga mass; and no Algoman granite—younger than Knife Lake slate—has been recognized (except as Tanton maps parts of a single granite mass as of two ages). We leave the problem as yet unsolved with the question: Is there here a slate older than the Knife Lake group, or an acidic intrusive younger than the Saganaga mass? On the basis of Gruner's study (1941, p. 1610) it is possible to suppose that the porphyry is part of the Knife Lake group, and *neither* earlier nor later.

Medial Precambrian Era and Rocks

Knife Lake Group

Name: The name Knife Lake slates was first applied to slates and graywackes which had their typical development on and near Knife Lake (R. 7 W.) along the international boundary (Alexander Winchell, 1888). They are interstratified with what was called Ogishke conglomerate and Agawa iron-bearing formation. Van Hise and Clements (1901) called them "Knife slates." By 1903 (Clements, 1903, p. 275–349) the name was used for belts of slate and graywacke from Vermilion Lake to Saganaga Lake. Correlation of these rocks with similar ones in Canada has been only tentative, and no formation seems to be equivalent on the south side of Lake Superior (Leith, Lund, and Leith, 1935, p. 17). As the Knife Lake slates and graywackes are definitely interstratified with what were once called Ogishke conglomerate and Agawa formation, Gruner calls them all the Knife Lake group or series, as did Grout (1933a, p. 992), and suggests that the two other names be dropped. As a name for the group, Knife Lake is better than Lower Huronian, since it was followed by great intrusions and erosion, and this important break would separate it from any Middle and Upper Huronian that might be near. To class the Knife Lake as Huronian would use the same name for two groups separated by an erosion interval of hundreds of million years. To call the Knife Lake pre-Huronian, as has been done, would either put it in the Keewatin, which ends with granite intrusions like the Saganaga granite—definitely older than the Knife Lake sediments—or leave it between the Keewatin and Huronian. The latter option is here preferred.

Areal distribution: The rocks of the Knife Lake group include from 10 to 20 recognizable members, well exposed along the international boundary between Vermilion Lake and Saganaga Lake, though exposures are scattered as far west as Koochiching County. The belt is about 70 miles long and 10 to 15 miles wide. Much has been mapped in great detail by Gruner, Dutton, and Gibson (Gruner, 1941). It has not been clearly identified east of the older Saganaga granite though a few small slate exposures are known. The Duluth gabbro and Giants Range granite, both intrusive into the Knife Lake series, form the southern border of the main belt. The northern and western limits of the area are rather indefinite because largely drift-covered, and because intrusive granites have metamorphosed the slates to mixed gneisses. There are scattered roof pendants in the granite. Separated from the type locality by several miles of intrusive granite are large areas of slate and schist that most geologists agree are well correlated with the Knife Lake group. The best known of these are in the so-called Virginia Horn between Gilbert and Virginia and farther westward.

There are other slates and schists, some well correlated with this group, but many miles of glacial drift-covered areas separate them from the Knife Lake area. Such are the Thomson formation near Cloquet and Carlton, Minnesota, and the great area near Rainy Lake where Lawson (1913) named the sediments the Seine series. These are rather clearly equivalent to the Knife Lake group (Grout, 1925a), and the latter name has priority. (*See* the pages on Thomson formation.)

Lithology: The rocks of the Knife Lake group include almost every kind of a metamorphosed mechanical sediment (Clements, 1903, p. 275–349). To think of them as dominantly slates would be wrong, for there are prominent conglomerates and graywackes, vol-

canic tuffs, and agglomerates (Gruner, 1941); and some flows and feeding dikes, erupted during sedimentation, are included. Near the larger intrusives heat has produced schists and hornfelses that reach a high grade of metamorphism. Almost all the flows and tuffs are andesitic, though some are rhyolitic. Most of the sediments seem to be derived from: (1) greenstones and andesitic extrusives or intrusives, (2) Saganaga and similar granites and gneisses, (3) rhyolite porphyries of about the age of the granites; locally there are scattered fragments of jaspilite.

Since very rapid changes in conditions of deposition suggest the sediments are continental, it is not surprising that conglomerates and agglomerates alternate with graywackes, slates, and tuffs. Small lenses of iron-bearing formation, formerly called Agawa formation by Clements (1903), appear interstratified with the tuffs, but some are replacements rather than chemical precipitates in shallow water (Gruner, 1941, p. 1616). Not all of the many members of the group are in continuous exposures.

Lithologically the conglomerates are most easily studied. They were formerly thought to be basal to the slates, and were called Ogishke conglomerate after particularly prominent outcrops along the north shore of Ogishkemuncie Lake. It is now known, however, that conglomerates occur at many different stratigraphic horizons in the series. Some lie directly on the older basement, in which case the pebbles and cobbles are made up largely of fragments of the rocks immediately below them. A conglomerate on greenstone flows consists largely if not entirely of greenstone fragments. Another on Saganaga granite is composed largely of granite fragments.

Since the granite pebble conglomerate, resting on the Saganaga granite, can be traced along the strike to places where it lies on slate, question may arise as to whether granite was anywhere exposed to erosion at the beginning of Knife Lake sedimentation. It might even be asked if the Saganaga granite could have been intruded after the lower members of Knife Lake group were deposited. Gruner (1941, p. 1635–1640) thinks the granite was intruded and the major deformation occurred before extensive deposition, but perhaps the granite was not exposed by erosion until some time after Knife Lake sedimentation began.

Light to dark-gray and greenish-gray chert fragments also occur in the conglomerate. Some of these contain remains of microscopic algae (Gruner, 1923, p. 147), thought to be of the same age as the Soudan jaspilite.

Conglomerates of greenstone fragments may resemble the original greenstone so closely that they have been mistaken for greenstone. Only after considerable weathering can the fragmental nature be recognized. The same problem arises in conglomerates of Saganaga granite fragments resting on the granite. Many students of geology have examined such outcrops on the west shore of Cache Bay of Lake Saganaga, and it requires very close study to determine their true character (Gruner, 1941, p. 1628, Pl. 4).

The conglomerates, across or along the strike, commonly grade into arkosites and graywackes. The arkosites are yellowish to green or red, and some have been mistaken for somewhat sheared granite. The graywackes are darker, mostly grayish green. They weather buff and brown like the greenstones.

The slates and the tuffs are of similar colors, gray predominating. They are well banded as a rule, and the flinty ones break with conchoidal fracture. Near the later intrusive granites the slates and graywackes become phyllites and schists. Near the later gabbro the slates become typical hornfelses. The deformed conglomerates commonly have fragments highly drawn out.

STRUCTURE OF THE KNIFE LAKE ROCKS: The Knife Lake sediments are highly folded and faulted. Only the area described by Gruner (1941) has been mapped in detail. Locally there are close folds in which the limbs and axial planes are nearly vertical. The folds in the major synclinoria generally plunge about 10° to 25° E. or W., but this is a generalization with many exceptions—detailed maps and sections (Gruner, 1941) should be consulted. Faulting is commonly parallel to the regional cleavage and schistosity, but is not observed in the field except after careful plotting and "matching" of all the folds. Faults are nearly always topographic "lows," covered by lakes or swamps. The displacements are large and may be miles horizontally though no large

ones could be measured. The faults cut the region into a number of segments, in each of which only a few beds can be mapped structurally. Beds in one segment cannot always be correlated satisfactorily with beds in another segment, because of a lack of knowledge of the displacements. Faulting is particularly common and difficult to interpret where it occurs between greenstone and fine-grained sediments (Gruner, 1941, p. 1622).

THICKNESS OF GROUP: The highly folded character of the rocks and their segmentation by faulting make estimates of the thickness of the sediments unreliable. The Knife Lake beds, however, where mapped in detail, have not been squeezed to such an extent that they are very much thinner than originally. If the deepest synclinorium representing the greatest accumulation of sediments is selected the following figures are derived for the total thickness (Gruner, 1941, p. 1624).

	Feet
Ester Lake graywackes, slates and tuffs	2000–4000
Kekekabic tuffs, agglomerates, and slates	2000–4000
Amoeba Lake graywackes, slates, and tuffs	4000–6000
Massive arkosites and graywackes	1000–2000
Well banded Knife Lake slates and graywackes	2000–4000
Basal gray slates and graywackes	500–1000
	11500–21000

The minimum 11,500 feet is probably closer than the maximum to the original thickness of the group. This particular stratigraphic column does not contain any andesite flows nor any prominent thick conglomerate, but has many smaller lenses, not individually named.

CONTACTS WITH OLDER ROCKS: Contacts between the Knife Lake and definitely older rocks are not common. Fault contacts are numerous between greenstones and the younger rocks, and, besides these, many contacts are concealed by depressions or dense undergrowth. Exposures for study are listed, but some of them are difficult of access, and not all can be described so that a visitor can easily find them. They are chosen from the area worked in detail near the international boundary east of Ely (Gruner, 1941), the only large area where the age relationship of the several formations and members has been clearly determined.

CONTACT WITH GREENSTONE: (1) The highest elevation of the area at Twin Peaks (sec. 4, T. 64 N., R. 6 W.) is a mass of greenstone almost surrounded by slates, graywackes, and greenstone-conglomerate. The conglomerate on the west side of the peaks in secs. 3 and 4, T. 64 N., R. 6 W., is in contact or nearly in contact with the greenstone at a number of places for about a mile. There can be no doubt of the relationship of the two rocks. The greenstone is ellipsoidal at many places, and the rounded fragments in the conglomerate are practically all greenstone or green andesite.

(2) On the northwest side of the same mass, in sec. 34, T. 65 N., R. 6 W., slate rests on the greenstone. At a number of exposures the contacts (except for 3 or 4 inches on the slate side) could be made visible by scraping off moss and earth with the hammer. There is no question that slate lies directly on greenstone, for gradational bedding only short distances away shows the tops of the beds away from the greenstone.

(3) An easily accessible place to study a contact is in E ½ sec. 24, T. 65 N., R. 7 W., at the west end of the largest of three islands in Knife Lake. Slate and greenstone are in contact on a cliff-like outcrop near the water's edge. The slate shows good gradational bedding with tops away from the greenstone. There is no conglomerate between slate and greenstone and no concealed belt or fault here. (A fault passes just south of this island through the lake from east to west.)

(4) South of Moose Lake in secs. 31, 32, 33, T. 64 N., R. 9 W., and secs. 4 and 5, T. 63 N., R. 9 W., a number of almost parallel synclines strike northwest-southeast. The anticlinal greenstone between the synclines is in contact with and overlain by greenstone conglomerate. The latter ranges from a few feet to 300 feet in thickness. It grades into graywacke and slate which show gradational bedding with their tops away from the greenstone. To reach these exposures one should leave the Winton-Moose Lake road near Moose Lake and walk along the contact eastward; or follow one of the parallel belts by walking from the shore of

Moose Lake southeastward; or make a traverse south from the portage a short distance west of Flask Lake.

(5) A granite pebble conglomerate lying on ellipsoidal greenstone in secs. 10 and 11, T. 65 N., R. 6 W. east of Hanson Lake is the only one containing many granite pebbles (along with greenstone and graywacke) seen in actual contact with ellipsoidal greenstone (Gruner, 1941, p. 1600). The conglomerate grades into graywackes and slates, in which gradational bedding proves that the sediments rest on the greenstone.

CONTACT WITH SAGANAGA GRANITE: The best-known outcrops of conglomerate consisting of granite boulders occur in Cache Bay of Lake Saganaga on the Canadian side of the international boundary. They are easily accessible by canoe or boat on the west side of the bay where several peninsulas project into the bay (Gruner, 1941, Pl. 1). The conglomerate at the very contact consists of boulders of granite which are easily recognized as the same granite as that on which the conglomerate rests. No bedding is visible exactly at the contacts, but as the conglomerate grades into arkosite it shows distinct bedding which dips westward at about 70°. Locally some cross-bedding shows that the tops of beds are away from the granite. Where the arkosite grades into graywacke slates, the gradational bedding also shows that the tops of the beds are to the west.

CONTACTS WITH YOUNGER ROCKS: The Knife Lake sediments are intruded by Giants Range granite (C. K. Leith, 1903, p. 85) northwest of Mesaba station and north of Mountain Iron. Some porphyries, possibly of similar age, cut the sediments between Ensign and Birch lakes (secs. 5 and 8, T. 64 N., R. 8 W.). The large mass of the (probably Algoman) Vermilion granite, in and around Basswood Lake, has metamorphosed the Knife Lake rocks profoundly to mica schists at the contact (Grout, 1925b, p. 473; 1933a, p. 1001) and to phyllites somewhat farther away. The smaller stocks of Snowbank Lake and Kekekabic Lake granites or syenites have had a correspondingly lesser effect but still sufficient to be plainly visible where contacts can be found (Balk and Grout, 1934, p. 625; Stark, 1927, p. 723).

The relationship of the Gunflint formation to the underlying rocks has been described by Clements (1903, p. 388). It can be studied with respect to the Knife Lake beds in secs. 25, 34, and 35, T. 65 N., R. 5 W., where it lies on the truncated andesite agglomerates and conglomerates. It forms an east-west belt with a dip of 30°–45° S. A large part of the contact is concealed by Chub River and Fay Lake which fill the depression along the contact. Farther east the iron-formation lies directly on greenstone (Fig. 8). Near Biwabik on the Mesabi range the Animikie lies diagonally across the contact of the greenstone and the Knife Lake group (C. K. Leith, 1903, p. 181). Along the railroad south of the Spring mine, sec. 11, T. 59 N., R. 14 W., an early railway cut exposed the basal Animikie conglomerate in a nearly horizontal bed, resting on the truncated vertical cleavage of Knife Lake slates (Grout and Broderick, 1919, Pl. XIII; Fig. 6).

The Duluth gabbro invades the Knife Lake group along the southern border. It is not clear whether the contact follows any structural weaknesses in the rocks of the group or pushed its way in along an old erosion surface of the Knife Lake rocks upon which Keweenawan rocks were deposited. The latter is more probable for it is the method of invasion on the Gunflint and East Mesabi ranges. In secs. 29 and 30, T. 65 N., R. 5 W., three outliers of gabbro extend almost a mile north of the main mass of the gabbro (Gruner, 1941, Pl. 1). They may be parts of a sill which invades the Knife Lake rocks along an almost horizontal contact. Possibly the lower contact of the gabbro follows an erosion surface. Metamorphism by the gabbro is not so extensive as by the granite masses. Nevertheless, the effect is plainly visible for hundreds of feet north of the contact. The Knife Lake rocks affected by the gabbro are dark and recrystallized to hornfels.[2] This contact metamorphism may be seen at many places. A good exposure lies north and northwest of Jordan and Ima Lakes, T. 64 N., Rs. 7 and 8 W. Another more easily reached is southwest of Snowbank Lake, on the old, long portage to the Kawishiwi River.

Trap dikes, some of which are probably

[2] Once called "muscovadite" from its resemblance to brown sugar (Grout, 1933a; N. H. Winchell, 1900, p. 67).

Keweenawan, cut the Knife Lake sediments in easily accessible places at Snowbank Lake, and in the east bay of Ensign Lake. Coarser dikes of diabase, also apparently Keweenawan, cut the "Seine" sediments of age equivalent to Knife Lake group, on Rainy Lake (Fig. 3). Cretaceous clays overlie the Thomson schists south of Little Falls (Grout, 1919, p. 198–199).

IS THERE JUSTIFICATION FOR THE NAME OGISHKE CONGLOMERATE? Clements (1903, p. 287–317) and his associates assigned a special position to the conglomerates in the Knife Lake group. They thought them basal to the other sediments and therefore their maps show the conglomerate outcrops as anticlines. They assigned the formation name Ogishke conglomerate to them presumably from the type locality on the north shore of Ogishkemuncie Lake. Actually the conglomerate is not basal at that locality but lies on slate (Stark and Sleight, 1939; Gruner, 1941, p. 1597). Near Alpine Lake (sec. 7, T. 65 N., R. 5 W.) the conglomerate lies on ellipsoidal greenstone and Saganaga granite, but if it is traced west it lies above other sediments that lie on greenstone (Gruner, 1941, map).

At several places slate and graywacke lie directly on the pre-Knife Lake rocks, and thick conglomerate lenses, even very coarse ones, may be interstratified with the other sediments without regard to the position of the basement complex. Evidently under these conditions either a different formation name would have to be used for each conglomerate lens or, as we prefer, the name Ogishke conglomerate should be dropped.

No particular localities need to be pointed out to convince one of the truth of the statement above. Almost any traverse through parts of the group will take one across several lenses or beds of conglomerate. An easily accessible syncline with two conglomerates, an agglomerate, and a slate lies about a quarter of a mile east of the northeast end of Moose Lake (T. 64 N., R. 9 W.).

ABSENCE OF A DEFINITE AGAWA IRON-BEARING FORMATION HORIZON: The "Agawa iron-formation," described especially by Clements (1903, p. 324) has really no claim to a formation name. Magnetite beds occur at several horizons in the Knife Lake group and, except on Agawa Lake, Ontario, are not more than a very few feet thick. Even the thin beds are not rich in iron. In the district under consideration and probably in the Canadian Shield in general, there was a tendency for deposition of ferruginous chert on a small scale through most of Knife Lake time. Almost anywhere in the group there may be small bands or lenses of chert, some containing iron oxide layers, or siderite, or ankerite.

Red jaspilite similar to Soudan formation but on a much smaller scale may be observed in secs. 2 and 3 T. 65 N., R. 6 W. (Gruner, 1941). This locality may be reached by portaging from Cypress Lake into Ester Lake and walking a short way southwest from the southwest end of that lake.

Other lenses of iron-formation have been described by Stark (1929) near Pickle and Spoon lakes in secs. 25, 26, and 35, T. 65 N., R. 7 W., reached by portages from Knife or Kekekabic lakes. Stark reports that at least part of the sediment consists of sands in which the iron-bearing mineral grains were mechanically concentrated.

Very crumpled jaspilite, a few inches to a few feet thick, may be seen on the south side of Neil Point on Rainy Lake. Lawson (1913, p. 66) noted that locally the iron-bearing beds occur where the basal conglomerate of the Knife Lake (Seine) sediments is missing.

Some of the "Agawa formation" of the older surveys (Clements, 1903, p. 329) is really a replacement. A striking example lies on the east shore of Ogishkemuncie Lake in secs. 23 and 24, T. 65 N., R. 6 W. It was supposed to be Agawa formation more than 1500 feet long and several hundred feet wide. In 1925 Gruner recognized it (1941, p. 1616) as a replacement in a shear zone of a big fault. The rock is made up of iron carbonate, chlorite, sericite and quartz and is weathered conspicuously to a chocolate brown along the lake shore and on several small islands. This zone actually cuts across the beds of the Agamok syncline in many places and replaces them by iron carbonate. As this replacement zone crops out on the shores of the lake it is easily accessible by canoe. A traverse (Gruner, 1941, Pl. I) southeastwardly from the fault a short distance is

sufficient to convince one that the "Agawa formation" here is a replacement.

Thomson Formation Problem

NAME: Thomson was the term introduced by Winchell (*See* Spurr, 1894a) for the slates exposed in Carlton County particularly at the village of Thomson on the east bank of the St. Louis River. They have also been referred to as St. Louis slates, Cloquet slates, and Carlton slates. Schwartz (1942b) suggested using "Thomson formation" since massive graywacke beds make up a considerable part of the formation and to the southwest metamorphism has produced phyllite and schist.

AREAL DISTRIBUTION: The main area of the exposures extends from 2 miles west of the Duluth city limits (Sec. 17, T. 49 N., R. 15 W.) southwesterly to the St. Louis River at Cloquet, Carlton, and Thomson. From Carlton scattered exposures occur in a belt a few miles wide, southwesterly to Denham (T. 45 N., R. 21 W.) in Pine County. At Denham a granite belt cuts off the outcrops and has metamorphosed adjacent portions of the formation to mica schist and mica-garnet schist. Far to the southwest at Little Falls (T. 129 N., R. 29 W.) on the Mississippi River are exposures of phyllite and schist which have long been correlated with the Thomson formation. Field work by Margaret Skillman (Mrs. Mark Woyski) in 1944 and 1945 has shown that several inclusions of metamorphosed Thomson formation occur in the granite between Denham and Little Falls, which may be recognized as Thomson by the characteristic concretions. There is now no reasonable doubt that the Thomson once existed as a continuous belt between Duluth and Little Falls, a straight-line distance of 140 miles. How far it extended beyond these limits can only be imagined. Near Duluth, for example, the slate disappears beneath basal Keweenawan.

LITHOLOGY: The Thomson formation consists of interbedded slates, graywacke-slates, and graywackes, varying southwesterly with the progressive metamorphism to phyllite, mica schist, and garnet-mica schist. Locally graywacke excludes other facies, and, between Cloquet and Scanlon, slate greatly predominates. The alternation of the three principal facies is characteristic. An excellent exposure of an 830-foot typical section along the St. Louis River in the gorge below the Thomson dam shows the following proportions: graywacke 48, graywacke-slate 25, slate 26. Although these proportions cannot be applied to the entire formation, they probably are representative. The detailed petrography has recently been described by Schwartz (1942b; 1949).

Graywacke: Graywacke beds range from a few inches to many feet thick. Most beds are relatively fine-grained, but some angular clastic grains can be recognized megascopically. The coarsest graywacke observed, in the north bank of the St. Louis River in sec. 13, T. 49 N., R. 17 W., has subangular grains up to 3 mm. in diameter. An exposure on Highway 61, 3 miles southwest of Carlton (sec. 16, T. 48 N., R. 17 W.), contains a few siliceous pebbles and concretions.

Graywacke-slate: A common rock interbedded in the Thomson formation is intermediate between slate and graywacke. It is more quartzose and has less flow cleavage than typical slate, although thin sections usually show orientation of the micaceous minerals. The minerals of the graywacke-slate and of the slate are quartz, feldspar, chlorite, and sericite, with small amounts of carbonate, pyrite, and leucoxene.

Carbonate concretions: Characteristic of the Thomson formation are the abundant carbonate concretions, described by Schwartz (1942c). (*See also* Harder and Johnston, 1918, Pl. X.) They occur throughout the formation and are useful in correlation. Many are near the tops, but some are scattered in massive beds. They are usually in zones parallel to the beds and are a valuable guide to the dip and strike. They range from less than an inch in diameter to fully 3 feet. They are most commonly roughly prolate ellipsoids, slightly flattened in the plane of the cleavage which passes through them. Most of the concretions consist mainly of calcium carbonate, but iron carbonate and probably manganese carbonate locally oxidize to yellow and black residues.

THICKNESS: Although the detailed structure of the Thomson slate cannot be worked out, available data on thickness suggest some limits.

The main area of the formation is exposed at intervals over an area 40 miles east and west and 25 miles north and south, throughout which east-west strikes of fold axes, beds, and cleavage are noteworthy. N. H. Winchell (1899b, Chapter 30) noted that prevailing dips are to the south. Tabulation of 544 dip observations by the writer shows that 71 per cent are to the south. North dips average 57°, and south dips 53°. Dips to the east or west are negligible. Graphic solution of the problem of thickness shows that 42 per cent of the distance across the strike is not opposed by an equal thickness dipping north. With allowance for dip there is an apparent thickness of 45,000 feet of beds in a north-south extent of 25 miles. Allowing a 50 per cent reduction for flat dips on the axes of anticlines and synclines and other causes of duplication, the estimated thickness is more than 20,000 feet.

STRUCTURE: Schwartz (1942a) described the structural details shown by the Thomson beds, and only the regional features pertinent to the suggested correlation are summarized here.

The location of the Thomson area west and southwest of Duluth places it well on the north limb of the Lake Superior geosyncline and about 10 miles north of the minor axis which passes south of Duluth and Superior. The formations of known Animikie and Keweenawan age dip about 10° SE. In contrast the exposed Thomson beds strike nearly east-west and dip north or south at angles averaging 55°. The great structural discordance between basal Keweenawan beds and the Thomson is well shown in the channel of the St. Louis River in sec. 16, T. 48 N., R. 16 W. The Thomson beds are folded, but only locally can the complete major structure be worked out because in many places no key beds can be seen, and outcrops are intermittent. The slaty beds have a well-developed flow cleavage which dips more steeply than the bedding, or is almost parallel to it (Harder and Johnston, 1918, Pl. VII, B). The nearly symmetrical open major folds strike nearly east-west. Isoclinal and overturned folds appear only near the granite contacts. The plunge of the folds is mainly 10° to 20° E., but gentle west plunges are not uncommon.

Fracture cleavage is abundant at places. Recognizable faults are not common, but in the absence of good horizon markers detection is difficult. Faults probably do not play an important part in the major structure.

CONTACTS: Nowhere is the base of the Thomson formation exposed, and there is little evidence of the formation on which it rests.

The Thomson formation was metamorphosed to a slate, deformed so that it stands vertically at places, then eroded before the Puckwunge conglomerate of the Lower Keweenawan was deposited on it. Good exposures lie near the west city limits of Duluth (secs. 17 and 20, T. 49 N., R. 15 W.) and along the St. Louis River in Jay Cooke State Park (Schwartz, 1949). The Puckwunge and overlying lowest lava flows of the Keweenawan dip gently southeast toward the Lake Superior syncline.

SIGNIFICANT EXPOSURES: In connection with the problem of correlation of the Thomson formation a few exposures are of much significance, but to understand the problem one must be familiar with the character and structure of the associated formations described elsewhere in this paper.

The series of exposures in sec. 20, T. 49 N., R. 15 W., are particularly important. A short distance south in sec. 15, T. 48 N., R. 16 W., in the St. Louis River channel, Puckwunge conglomerate lies on steeply dipping Thomson slate (Fig. 9).

In sec. 1, T. 48 N., R. 16 W., on Little River near its mouth on the St. Louis River, slate, Puckwunge conglomerate, and upper (or middle) Keweenawan sandstone are exposed within a few yards of each other.

The general nature of the Thomson formation may be easily seen in the village of Carlton and along the St. Louis River from Cloquet through Jay Cooke State Park to Fond du Lac, and more metamorphosed facies may be seen along the Soo Line railroad tracks northeast of the village of Denham (T. 45 N., R. 21 W.) in Pine County. Metamorphosed concretions are well exposed near Little Falls on the Mississippi River, where a creek enters from the west, about 2 miles north of the highway bridge.

CORRELATION: The slates in the Carlton-Cloquet area have been variously correlated since 1883. To avoid a long discussion, the srevious suggestions are given in tabular form.

Author	*Date*	*Correlation*
Irving	1883	Animikie
Winchell, N. H.	1893	Animikie and Taconic (2 series)
Spurr	1894	Keewatin (Lower Huronian)
Winchell	1899b	Animikie and Keewatin (Lower Huronian) (2 series)
Leith	1903	Lower Huronian
Van Hise and Leith	1911	Animikie (Virginia slate)
Harder and Johnston	1918	Animikie (possibly Lower Huronian)
Leith, Lund, and Leith	1935	Upper Huronian (Animikie)
Schwartz	1942b	Pre-Animikie

Most of these correlations were made either on broad general grounds or purely on lithologic characteristics. Correlations previous to that by Schwartz (1942b) were made without all the facts now available, and it is believed that a more definite conclusion is now possible, although it may be that lack of continuous exposures precludes exact correlation.

The following facts must be taken into account:

(1) The Thomson formation differs lithologically from the Animikie sedimentary rocks of the Lake Superior district (except perhaps some facies of the Rove formation) in having exceptionally abundant carbonate concretions and a high percentage of graywacke beds.

(2) The Thomson formation is highly folded and metamorphosed so that the lowest grade has excellent slaty cleavage. In this respect it resembles the Knife Lake series of the Vermilion district and not the Animikie Virginia slate. The Cuyuna slate series has such cleavage, but its age is still open to discussion.

(3) The strikes of the beds, cleavage, and axial planes of the folds are predominantly east-west, at an angle of about 35° to the prevailing regional structure of younger beds related to the north limb of the Lake Superior syncline. At Duluth the two structures strike about at right angles to each other.

(4) The Thomson formation is separated from the lower Keweenawan by a pronounced angular unconformity, whereas the Animikie in the Grand Portage area on Lake Superior lies almost conformably under Lower Keweenawan.

(5) A pronounced structural discordance exists between the Animikie of the Mesabi district and the Thomson formation. The two are not exposed in contact, but both the Mesabi formations and the Keweenawan of the north shore of Lake Superior partake of the regional structure of the Lake Superior syncline. As previously shown the structure of the Thomson formation is completely at variance with this regional structure (Schwartz, 1942a, 1949).

(6) The Thomson has suffered progressive regional metamorphism with low-grade rocks in the Carlton-Cloquet area and high-grade metamorphics farther southwest near the granite.

(7) On the basis of accessory-mineral studies the granites of central Minnesota have been tentatively correlated (Tyler *et al.*, 1940) with granites of Algoman age. Seven samples of granites collected in 1941 from scattered points from Little Falls to Denham show accessory minerals characteristic of Algoman and older granites. The only batholithic Minnesota granite which shows Keweenawan characteristics is the series of outcrops in the Milaca and St. Cloud areas far to the south (Woyski, 1949).

There is little direct evidence as to the correlation of the Thomson formation with other pre-Animikie rocks. On the basis of lithology, metamorphism, and deformation, the formation compares well with some of the Knife Lake group, and reference to the recent summary of Lake Superior correlations (Leith, Lund, and Leith, 1935, p. 10) leaves no other logical choice. (*See also* Gruner, 1941.) This is not a new idea but was suggested by Spurr (1894), Leith (1903), and others. Regional correlation is difficult because the position of the Knife Lake group in the medial Precambrian is somewhat doubtful, and each group is complicated in itself.

In any event the Thomson formation seems to be pre-Animikie and pre-Algoman. Its character and metamorphism correspond well with what might be expected in a pre-Animikie formation.

Algoman Intrusives

The name Algoman was first applied by Lawson (1913, p. 5) to rocks in the old western Ontario district of Algoma. He based the name

on the fact that some granites are older and some younger than a well-marked sedimentary series, the Seine, now considered equivalent to the Knife Lake group. "Having restricted Laurentian to the older we recognize the need of a name for similar intrusives that are post-Seine and refer to them and to the period in which they were developed as Algoman." The granite which most clearly cuts the Knife Lake group and yet is older than the next recognized sediments, the Animikie, is the Giants Range granite cited by Lawson (1914). Many other masses are probably of the same age but are not exposed in contact with the two successive sedimentary groups.

Tanton (1927) suggested that intrusives *into the Huronian* may be called Algoman. On the other hand, Pettijohn (1937), having evidence of more than two early periods of *granitic invasion*, avoids the name Algoman. Leith, Lund, and Leith use the name Algoman for the Giants Range granite, since it is at the type locality and the relations fit the original definition; but they note that, since the Knife Lake group cannot be clearly determined as equivalent to any one of the Huronian divisions, the Algoman mass may be pre-Huronian, or formed during some part of Huronian time.

The Giants Range granite is a batholith about 100 miles long and up to 15 miles wide. It has several facies, and they have been mapped and their sequence investigated (Allison, 1925). The petrographically oldest-looking material north of Mesaba is gneissic and may be a Laurentian inclusion, but contacts seem gradational. The east half of the batholith is mostly hornblendic granite and older than the biotite granite of the west half. Inclusions of large blocks of hornblende granite in biotite granite may be seen north of Virginia and north of Mountain Iron (T. 59 N., Rs. 17 and 18 W.). Both facies grade locally along the borders to darker rocks—syenites, shonkinites, and hornblendites. There are small monzonitic and even gabbroic facies. The zircons of Algoman granites (Tyler *et al.*, 1940) are indistinguishable from those of the Saganaga granite—they are of the hyacinth type.

On the map of the Mesabi district (Van Hise and Leith, 1911) one small area of granite near Embarrass, Minnesota, was mapped as intrusive into the Animikie iron-formation at the east end of the range. Even as late as 1945, it was claimed (Royce, 1945) that a later granite intrudes the iron-formation at the east. Grout and Broderick (1919) had the advantage of studying the exposures on the east Mesabi just after a forest fire had cleaned off many exposures that have been concealed before and since by a dense growth of moss and brush. The relations seemed very definite and may be summarized as follows:

A few granitic stringers cut the iron-formation from Babbitt northeast, becoming more common close to the Duluth gabbro (T. 60 N., R. 12 W.). Few of them are more than an inch wide, and the definite association with gabbro leads to the positive opinion that they are pegmatitic emanations from the gabbro into its floor. Similar pegmatitic stringers occur in the slates for many miles along the gabbro floor, but no large mass of related granite has been discovered.

The granite north of the Animikie belt near Embarrass is unconformably below and older than the Animikie group. Figure 5 is typical of the contact relations all along the east Mesabi range. It is noteworthy that some basal conglomerate rests on rather steep slopes along the sides of pre-Animikie valleys. Where these contacts are not well exposed some observers may be led (Rickarz, 1930) to think that the granite transgressed the gently dipping beds, but the many good exposures showed no such transgression. Another feature that might be misinterpreted is the increased metamorphism at the east end of the range. A reconnaissance study might attribute the metamorphism to the heat of a granite north of the range as logically as to that of the gabbro south of the range. Detailed study makes it very clear that the gabbro supplied the heat, even recrystallizing the granite where the base of the iron-formation lay on the granite. An outcrop of granite near the later gabbro in the NW¼ sec. 3, T. 62 N., R. 10 W., is recrystallized by gabbro heat so that its texture is notably changed (Grout, 1937, Pl. 5, fig. 2).

At some mines, between the east end and the central part of the range, dikelike masses seem to cut the ore. They are much altered, and their original character is obscure. Almost certainly however, some igneous intrusives are later than the iron-formation, but no large body of granite

has been found along the Mesabi sending dikes into the formation.

The Giants Range granite intrudes the Knife Lake group at many places (Leith, 1903, p. 84–85). Northwest of Mesaba station in the SE¼ of SE¼ sec. 18, T. 59 N., R. 14 W., and northeast of Mesaba station 50 paces north of the east quarter post of sec. 9, T. 59 N., R. 14 W., contacts are well exposed. The graywacke slates with thick to thin beds are shot through with stringers of granite, some parallel to the structure and some transecting it. Other even more complex contacts are exposed (1) north of Mountain Iron, northwest of the center of sec. 34, T. 59 N., R. 18 W., and (2) northeast of Virginia in sec. 34–35, T. 59 N., R. 17 W.

Hinsdale siding along the railroad 2 miles north of Mesaba station shows characteristic fresh hornblende granite in an abandoned quarry.

On the north side of the Giants Range along the railroad from Embarrass north toward Tower (Ts. 60 and 61 N., R. 15 W.), outcrops and railroad cuts supply an instructive series. At Embarrass the granite is pink and porphyritic. About a mile north there are outcrops of mica schist, and the granite, much whiter, injects the schist *lit-par-lit*. In the next 2 miles the proportion of granite becomes progressively less, and the schist grades into graywacke and slate with alternating beds very characteristic of the Knife Lake group. This is part of a large area of slate reaching across into the Vermilion range and correlated definitely with the exposures on Knife Lake. Seven miles north of Embarrass the slate has local conglomerate beds. The Keewatin greenstone occurs along the railroad about 10 miles from Embarrass.

Farther east, the north side of the granite intrudes Knife Lake slates at many outcrops, some 7 to 12 miles east of Ely, but these are perhaps less easily accessible than those already mentioned.

North of the Giants Range across a narrow belt of schist in an area some 30 by 80 miles lies the Vermilion granite batholith, which closely resembles the Giants Range granite in its several facies. This is probably not conclusive as to its age, because the resemblance is not much greater than that of both granites to the earlier Saganaga mass—evidently petrographic features are not a safe basis for correlation. The Vermilion granite intrudes the Knife Lake at hundreds of places and is clearly younger, but how much is not clear. It is nowhere in contact with Animikie sediments to prove that it is older than Animikie. The zircons are nearly all of hyacinth type—pre-Keweenawan—so that the mass is generally thought to be Algoman. The magnetite associated with the Vermilion granite pegmatites gives a tentative age (Hurley, 1943) of 750 million years. The intrusion of the several facies was distributed over a considerable period (Grout, 1926, p. 31).

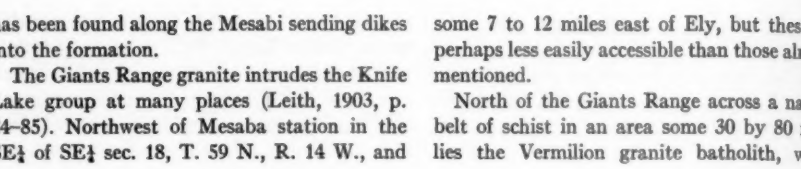

FIGURE 5.—SKETCH OF THE IRREGULAR GRANITE SURFACE ON WHICH ANIMIKIE SEDIMENTS WERE DEPOSITED ON THE EASTERN MESABI RANGE

North of the Vermilion batholith are some stocks on Rainy Lake. Several may be of the same age as the Vermilion, but not all are so similar petrographically to the Giants Range that the age can be assumed without question. On Grassy Island, Rainy Lake, (T. 71 N., R. 22 W.), there is not only a fresh-looking granite, but an older-looking belt of granite, and Lawson assigned one to the Algoman and one to the Laurentian (Fig. 3). At Lake of the Woods (T. 163 N., R. 34 W.) the granites somewhat resemble Algoman masses, but none have contacts to prove their age.

East of the Giants Range are some smaller intrusives, the Snowbank stock (T. 63 N., R. 9 W.) and Kekequabic stock (T. 64 N., R. 7 W.) which are probably Algoman, but they are

nowhere in contact with Animikie beds to prove their pre-Animikie age. They intrude the Knife Lake group at many places.

Many other granites in the State are probably Algoman, but their age cannot be settled by their relation to surface rocks of known age. Several have been tested for zircons, and since these are of hyacinth type they suggest a pre-Animikie granite. The series of granites and diorites from McGrath to Little Falls (T. 44 N., R. 23 W., to T. 40 N., R. 32 W.) intrude the Thomson slate (schist) that resembles Knife Lake slate much more closely than it does the Virginia slate. If any Minnesota granite of pre-Animikie age intrudes a slate, it is logical to correlate it with the Algoman granite intruding Knife Lake slate.

South of these intrusives from McGrath to Little Falls there are granites near St. Cloud (T. 124 N., R. 29 W.) that carry zircons of Keweenawan type, younger than Algoman (Woyski, 1949). Still farther south and west are granites in the Minnesota River Valley, which may be Algoman, but there is very little evidence as to their age.

The Algoman granites, younger than the Knife Lake group, naturally intrude also the older Keewatin greenstone and Soudan iron-formation. Outcrops may be seen along the highway near Armstrong Lake (T. 62 N., R. 14 W.). Others are mentioned in the description of the Soudan formation.

The widespread igneous intrusive activity in Algoman time shows that it was an important period of orogeny. High mountains were formed with complexly folded structures, involving both the Keewatin and the thick Knife Lake groups.

Post Algoman Unconformity

The unconformity below the Animikie cuts deep into the Algoman granite, though the Knife Lake group must have been thick in the mountain folds above the granite batholiths. Similarly in the absence of Algoman granite, where the Knife Lake rocks were deposited on pre-Knife Lake granite and Keewatin greenstone, the unconformity cut deeply through the slates and into the Saganaga granite and the greenstone.

The Animikie rests on Algoman granite at the village of Babbitt (T. 60 N., R. 12 W.) on the east Mesabi range. It rests on Knife Lake slate along the railroad near Spring Mine (T. 59 N., R. 14 W.) east of Mesaba; also farther east in less accessible outcrops on the south side of Disappointment Lake and north of French (Kakigo) Lake, transgressing the folded beds of different parts of the group. Animikie rocks (Gunflint formation) rest on Saganaga granite in the hills east of the Gunflint Trail (road) between Gunflint and Saganaga Lake (secs. 22 and 23, T. 65 N., R. 4 W.). Excellent exposures occur also on the Canadian side, a few rods north of "Bridge Bay" of North Lake. The same Gunflint belt lies on Keewatin greenstone near the Forest Service look-out tower (T. 65 N., R. 4 W.) some 3 miles west of Gunflint Lake. (*See also* the notes on Pokegama quartzite.)

Later Precambrian Era and Rocks

Animikie Group

General discussion.—Animikie is an Indian term for Thunder Bay, a few miles northeast of Minnesota, on the north shore of Lake Superior, where the rocks of this group are well exposed. Parts of the group in Thunder Bay can be related without much doubt to exposures in Minnesota and are fairly well correlated with the iron-bearing rocks across to Wisconsin and Michigan. The name was first applied by Hunt (1873) to beds underlying the Keweenawan and "overlying Huronian schists." Van Hise (1892) believed the group was Upper Huronian, overlying lower Huronian. The International Committee (Hays *et al.*, 1905) agreed. Having found an unconformity on the south side of Lake Superior, Allen (1919) tried to divide the Animikie into Middle and Upper Huronian, putting the iron-bearing rocks into the Middle Huronian, and an overlying slaty group in the Upper. Lawson (1929) concluded the Animikie is not Huronian in any sense. Andrew Leith (1935) recommended abandonment of the term Huronian because on the south shore it includes two unconformable series.

Tanton (1931) made a detailed study of the type locality and reports the Animikie lies

unconformably on the Earlier Precambrian schists and granite and disconformably below the Sibley—a group underlying the Keweenawan. We retain the name because Minnesota outcrops are well correlated with the original Animikie locality and not well correlated with the type locality of the Huronian.[3] Even if the equivalent beds south of Lake Superior make up two disconformable or unconformable formations, which are conformable in Minnesota, the term Animikie can serve for a group of formations.

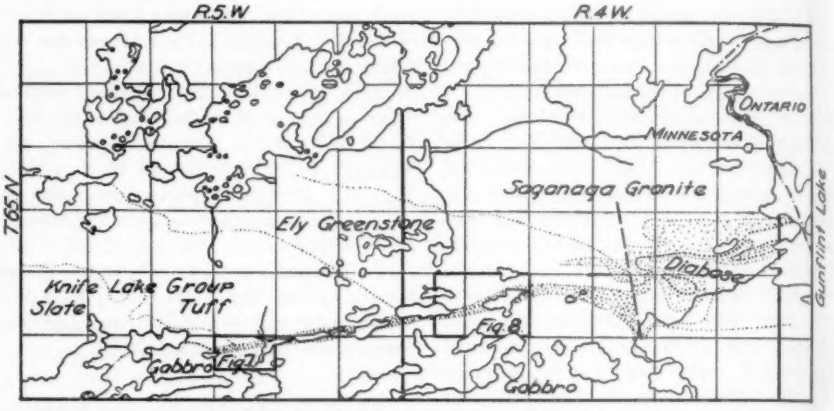

FIGURE 6.—KEY TO LOCATIONS OF FIGURES 7 AND 8

EXPOSURES OF INSTRUCTIVE CONTACTS: The unconformity of granite below Animikie is well seen at several places on the south slope of the high ground at Babbitt, secs. 17 and 18, T. 60 N., R. 12 W. The rough pre-Animikie topography can be seen along a sewer ditch southeast of the houses for the workers at the mill.

The granite of the type locality of Algoman is cut by trap dikes supposedly Keweenawan, along the road north of Virginia (T. 59 N., R. 17 W.).

Exposures showing the relations to the gabbro and to older formations have already been cited.

Pokegama quartzite.—The Pokegama quartzite was named after its prominent outcrops at Pokegama Falls on the Mississippi River north of Grand Rapids, Minnesota. The name seems to have been used first by H. V. Winchell (1893). Previous to this date the quartzite was usually referred to as a part of the Pewabic quartzite which then included most of the known quartzites of Minnesota according to the early geologists. It was recognized as the base of the Animikie group.

AREAL DISTRIBUTION: Pokegama quartzite lies on the truncated pre-Animikie rocks; that is, on Archean greenstone, on pre-Knife Lake granite, on Knife Lake group, and on Giants Range granite; and is overlain conformably, or nearly so, by the Biwabik iron-bearing formation. It either crops out or is found in drill cores along most of the Mesabi range. Toward the west end beyond Lake Pokegama drilling has not found its limits. Its strike parallels the Mesabi range. It dips to the south or southeast with the exception of the part between Virginia and Eveleth, in the so-called Virginia Horn. Dip angles range from 3° to 12°. Eastward it thins to a trace of conglomerate near Babbitt in T. 60 N., R. 13 W. (Fig. 5), and is very thin where the Animikie crops out near Gunflint Lake. Near Gunflint Lake the Animikie (with only thin basal quartzite and conglomerate) transgresses from the coarse tuffs of the Knife Lake group, onto the Ely greenstone and Saganaga granite (Figs. 6, 7, 8).

[3] Royce (1925) claimed that the United States and Canadian Surveys had united in restricting the term Animikie to the Upper Huronian, but we find no reference to such an agreement.

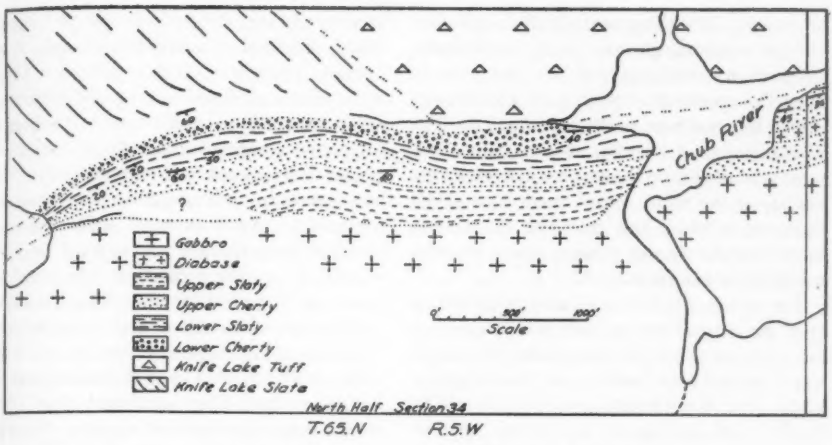

FIGURE 7.—ANIMIKIE SEDIMENTS LYING ON TWO FACIES OF KNIFE LAKE SEDIMENTS

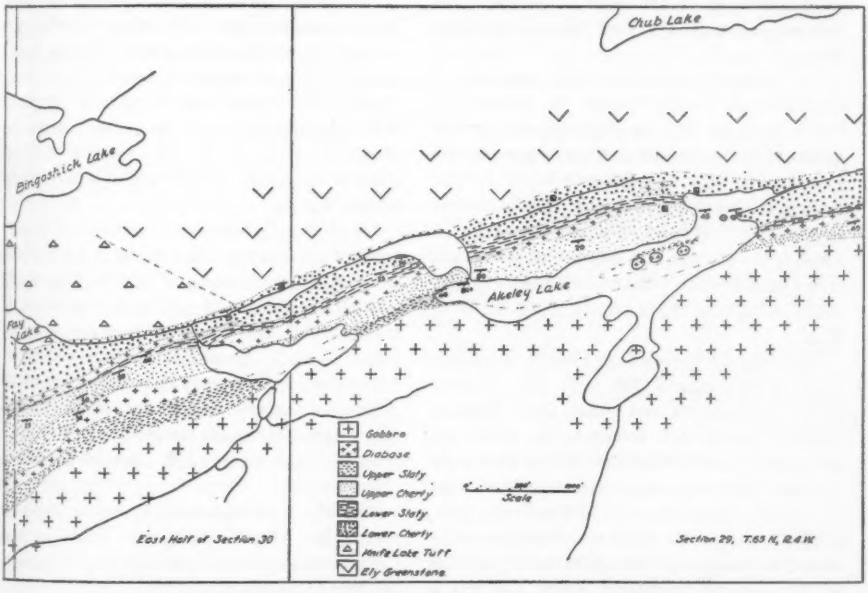

FIGURE 8.—ANIMIKIE SEDIMENTS LYING ON GREENSTONE

LITHOLOGY, STRUCTURE, AND THICKNESS: Across the strike the quartzite changes from a conglomerate at the base through thin-bedded micaceous quartzite to very massive coarse-grained quartzite at the top. This succession is fairly persistent where the quartzite is thick. The three rocks may differ widely, however, in thickness. The conglomerate consists largely of fragments of the rock immediately below it. The old erosion surface on which deposition occurred was rough and full of large cracks and fissures, some 6 feet deep, all of which were filled by usually angular fragments especially where the underlying rock consists of slate and

graywacke. Other fragments in the conglomerate are andesite, granite, quartz, and chert. Greenstone should supply pebbles but looks so much like graywacke in small weathered pebbles that it has not been positively identified. The sizes range from boulders to grit. There is a rapid gradation from the coarsest at the bottom. Insofar as the writer knows there is no single exposure in which this gradation can be observed completely, but at many places the very coarse facies can be studied.

The thinly bedded micaceous quartzite is gray, fine- to medium-grained, and breaks easily into slabs along the bedding planes. Jointing is nearly normal to the bedding and closely spaced in the more thinly bedded material. The thinly bedded rock is thicker in the belt east of Virginia than west of it. But few drilling data are available on it. There are only man-made exposures of it on the main range, and only thin exposures on the East Mesabi and Gunflint.

The massive quartzite at the top may be conglomeratic locally though the pebbles are rarely over an inch in diameter and are all quartz. This quartzite is pinkish to light greenish gray and is very massive and hard. It may become darker gray downward. This facies may be 40 to 50 feet thick. On account of its resistance it outcrops in a number of places and it has been used for crushed rock from a quarry north of Hibbing (S½ of sec. 25, T. 58 N., R. 21 W.).

The total thickness of the Pokegama quartzite does not exceed 200 feet, and at most places is probably not much over 75 feet, which is the greatest thickness the writer has seen in drill cores. On the East Mesabi Range the quartzite thins to a conglomerate a few inches to a few feet thick (Grout and Broderick, 1919, p. 18). It is not clear whether this conglomerate should be assigned to the quartzite or the base of the Biwabik formation which also has a basal conglomerate bed over the whole length of the range. There is nothing in the composition or position of it which could not belong to either one of the conglomerates.

CONTACTS AND SIGNIFICANT EXPOSURES: There are many exposures of the contact of the basal conglomerate and the truncated pre-Animikie rocks. Evidently this pre-Animikie erosion surface was at many places approximately like that of today. For example, in the Virginia Horn area, from a mile east of the main street of Eveleth to a hill almost 2½ miles north of it, patches of the basal Pokegama conglomerate occur on the Knife Lake group (slates, graywackes, and conglomerates). The best exposures are along the fringes of the pastures north and northeast of Eveleth where the first outcrops are encountered near the wooded areas. The ancient erosion surface exposed now is up to 1500 feet wide, and its slope corresponds to the dip of the Animikie series. These conglomerate patches may be only a few inches thick but filled some deeper cracks in the ancient surface. They are fairly easily recognized because the slaty and schistose fragments have random positions and are not lined up with the regional schistosity of the rocks underneath. As the pre-Animikie rocks contain thin beds of conglomerate with almost vertical dip one can readily find here gently dipping Pokegama conglomerate resting on steeply inclined conglomerates hundreds of millions of years older. Many similar outcrops occur north of the Mountain Iron pit (at Mountain Iron) just south of the height of land which is part of the Giants Range.

Another easily accessible area of such outcrops of basal Pokegama is north of the Lavinia location which is north of the Hull-Rust pit near Hibbing. A gravel road leads west of north from Lavinia past a quarry in massive Pokegama quartzite. About 500 to 600 feet north of the quarry, outcrops may be found on the south slope of the gentle rise of ground. Patches of very coarse Pokegama conglomerate lie on an eroded surface of graywacke intruded by Giants range granite.

Possibly some quartzites exposed at the open pits of the Maroco and Rowe mines on the Cuyuna range are equivalent to the Pokegama, but this has been questioned.

Biwabik iron-bearing formation.—Biwabik is an Ojibway word for a "fragment of iron," and for this reason and because the Biwabik Mine was one of the good early mines the iron-formation of the Mesabi range (formerly classed as Taconic, from which the rock name taconite is derived) was named the Biwabik formation by Van Hise and Leith (1901). They considered

it conformable with the Pokegama below and the Virginia slate above. C. K. Leith (1903, p. 167) refers to a "minor erosion interval" between the Pokegama and Biwabik, but that interval is commonly ignored. The conformity with the Virginia was not questioned until Allen (1919) reported an unconformity in districts south of Lake Superior, at about the same horizon. Since then several attempts have been made to separate the Biwabik and Virginia. The facts for Minnesota are given in detail.

AREAL DISTRIBUTION: The Biwabik formation in general has the same distribution as the underlying quartzite. It has been explored by drilling over a distance of about 100 miles on the Mesabi, besides about 50 miles on the Gunflint range, and possibly many more on the Cuyuna. Several of the magnetic belts west of the Cuyuna may indicate Biwabik formation, but only a few have been explored. The Biwabik is parallel to the granite ridges of the Mesabi Range to the north and may have extended north of the range at an early period though no evidence of it has ever been discovered. Toward the south it is overlain by the Virginia slate. Before the iron mines were opened up there were less than half a dozen natural exposures of the iron-formation west of the location which is called Mesaba Station today. East of it more outcrops occur so that the iron-formation on the East Mesabi Range was known almost 30 years before that of the main range.

LITHOLOGY AND INTERNAL STRUCTURE: The detailed mineralogy and geology of the Biwabik have been described recently by Gruner (1946). Most of the rocks are widely known as ferruginous cherts, as in other districts, but a rock name more restricted to this district is "taconite." First used by Winchell about 1890, it implied an age relation to the Taconic rocks in eastern States, with which Winchell thought they could be correlated. Taconite now applies to nearly all the rocks of the Biwabik formation (except the ore and possibly the intermediate slate). Prefixes show the very considerable differences in the formation—greenalite, cherty, slaty, granule, mottled, conglomerate, and many others.

The most striking peculiarity of much of the taconite is a granule texture, based on granules of about the size of a pin head (locally on the East Mesabi they are coarser and grade to pebble sizes). The granules were early noted by Spurr (1894b), later by Leith (1903). They differ from sand grains in structure as well as in form, though the differences are not so easily seen with the naked eye as under a microscope. At a few places and in certain beds there are oölites, but many more granules are massive or show only shrinkage cracks. Characteristic granules occur mostly in the cherty layers where beds are half an inch to several inches thick, but even the thin-bedded rocks may have granules. The most slaty layers have least. Any mineral of the taconite may be in granule form or may be the matrix to granules of the same or other minerals.

Certain major varieties of taconite serve as the basis of a stratigraphic sequence of divisions of the Biwabik formation into members. Wolff (1917) made the suggestion most commonly used on the range. It has been elaborated by some details and horizon markers. Gruner's (1946) most recent table is as follows:

SUBDIVISIONS AND HORIZON MARKERS IN THE BIWABIK FORMATION (AFTER GRUNER, 1946)

	Thickness in feet
Upper Slaty Division	
Limy carbonate with silicates and slaty taconite	0–25
Slaty and thin-bedded cherty taconite, silicate taconite. (On the East Mesabi; white quartz septaria in concretions; some beds rich in magnetite)	0–145
Upper Cherty Division	
Cherty thick-bedded, silicate taconite with relatively few thin beds (more and thicker at the east)	80–250
An algal horizon near the middle of the division is from 6 to 60 inches thick.	
Lower Slaty Division	
Slaty or thin-bedded taconite, silicate taconite; relatively few thicker cherty beds, considerable carbonate and thin conglomerate textures near the top. On the East Mesabi,—recrystallized	0–250
"Intermediate Slate", black and thin-bedded silicate taconite; paint rock	1–40

Lower Cherty Division

Lean member, white thick-bedded cherty taconite	10–50
Magnetite-rich member, banded, mottled, silicate and chert (On East Mesabi very thin)	2–250
Cherty thick-bedded members with ferric oxide, cherty and banded taconite, thin-bedded at top (Rarely occurs on East Mesabi)	8–70
Red basal taconite (absent on East Mesabi)	0–40
Basal conglomerate and coarse algal chert	0–15
Total	350–755

Gruner's (1946) chart of measured sections shows that these divisions and horizons can be traced along the Mesabi range. The change from Lower Slaty to Upper Cherty divisions is rather gradational and not easy to locate accurately.

The two important beds are the "algal" horizons. They range from a few inches to a few feet thick, and structures resemble contorted bedding. Horizontal sections show concentric banding like concretions. The forms near the base of the Biwabik are 1–2 feet across, but the zone or bed in the Upper Cherty division, especially on the East Mesabi, has structures like a pile of thimbles, in a series of columns, like fingers held upright. The algal beds are commonly red and white chert, but high-temperature metamorphism turns some of them gray. Where granules of taconite occur between the fingerlike algal forms, the bands merge into concentric layers around the granules, making an oölitic pattern.

The algal beds resist weathering and yield good horizon markers in some of the open pits, from Hibbing east, even through the Gunflint area.

THICKNESS: The iron-formation is 350 to 750 feet thick, thickest in the central part of the range, and thins gradually to the east and west. It has often been assumed that it continues far beyond its known extent to the southwest, but there is no conclusive geological reason why the formation could not pinch out. The iron-formation of the Cuyuna range may be such a continuation, as a number of geologists believe, but the evidence is not conclusive. Eastward the Gunflint formation is easily correlated with the Biwabik, though for about 40 miles between the two the formation has been all but obliterated by the invasion of the Duluth gabbro.

RELATIONS: For Minnesota areas *see* section on its relation to Virginia slate. As the iron-formation dips generally south at an angle of 3° to 12° and more, it has been thought probable that it passes under Lake Superior and connects with the Ironwood formation of the Gogebic range. It is very unlikely that we will ever know whether the connection exists for the observed dips would carry the formation to depths of 10 to 20 miles; but the correlation is good, stratigraphically, including sequence and even algal horizons.

CONTACTS AND SIGNIFICANT EXPOSURES: Very few natural outcrops of the iron-formation exist except on the East Mesabi range. The only easily studied one is a cliff of great prominence less than a mile north of Eveleth west of the main highway leading to Virginia. The top of the cliff is about 150 feet above the base of the formation. It consists of magnetic "irregularly banded" taconite of the Lower Cherty division. Another exposure is along the highway about halfway between Gilbert and Elcor. In constructing the highway an originally small outcrop of taconite was made into a large exposure of highly magnetic rock of the same horizon as the top of the cliff north of Eveleth.

The open pits of the range show many large artificial exposures of taconite, which, when highly oxidized, resembles high-grade ore. At a few mines where this oxidized material has been removed, unoxidized taconite is well exposed. The accessibility of such places changes from year to year depending upon minimg operations and flooding of the open pits.

The contact of the Biwabik formation and the underlying quartzite has to be studied largely in drill cores. As a rule the quartzite is massive at the contact. The first indications of iron-formation commonly consist of highly cherty coarse algal structures and pebbly beds of conglomerate. They are not necessarily in the order (1) conglomerate and (2) algal structures, but may be reversed so that algal structures are in contact with quartzite. The conglomerate is rarely over a few feet thick. The algal-structure beds may be 6 inches to 6 feet thick. No one

seems to think that an unconformity exists between the quartzite and iron-formation though the lithologic break between the two is very sharp, and the presence of conglomerate consisting of quartzite and chert pebbles attests to some erosion during the change from clastic to chemical sedimentation.

The actual contact between the quartzite and the iron-formation was exposed in 1950 at the extreme south end of the Minnewas open pit near Virginia.

Virginia slate.—The Virginia slate is named from the city of Virginia in the central part of the Mesabi Range. It has been found in numerous test pits and drill holes. It was first described as conformably above the Biwabik formation and below the Keweenawan Duluth gabbro.

AREAL DISTRIBUTION: The Virginia slate lies on the Biwabik formation, but questions have been raised as to whether the relation is conformable. The slate underlies a large area from east to west parallel to the Mesabi range and extends far southward from the range. The Rove slate is the equivalent of the Virginia slate on the Gunflint range. If the Cuyuna is eventually correlated with the Mesabi, some of the Cuyuna slates may be equivalent to Virginia slate.

LITHOLOGY AND STRUCTURE: The Virginia slate is not a slate but an argillite and graywacke. It has no slaty cleavage on the Mesabi Range as it was not dynamically metamorphosed. Much of it has conspicuous thin bedding, but the graywacke portions are thick-bedded. Our knowledge regarding this rock is based largely on drill-core information. Lately it has been exposed on the south side of the Embarrass Lake Mine near Aurora (T. 58 N., Rs. 15 and 16 W.) and in the Miller Mine at Aurora. There is no doubt that it lies conformably on the Upper Slaty division of the Biwabik formation at both places.

The Virginia slate is medium to very dark gray in most drill cores. Some portions are black, and these are usually thin-bedded. The formation has the same general dip as the Biwabik formation, commonly about 4°–30° S. Grant (1899) noted that the lower part of the Rove slate, Gunflint-range equivalent of the Virginia slate, was black and graphitic, and the higher horizons more sandy, grading from graywacke to quartzite, and locally carrying abundant concretions (Grout and Schwartz, 1933, p. 20). This is now considered gradational rather than the basis of a division of the formation. Deep drill holes south of the Biwabik outcrops show similar gray to black argillite and graywacke beds.

THICKNESS: The thickness has been estimated at several thousand feet (drill holes near the range cut 1800 feet). As seen in the cores the dip of the formation is between 4° and 10°. In Cook County the thickness may be 2000 to 3000 feet, but on the western Mesabi there may be much greater thicknesses. Royce (1945, p. 4) estimates possibly 40,000 feet but does not mention the places where the upper contact and overlying formations are known, as they are in Cook County.

RELATION TO OVERLYING FORMATIONS: Along most of the Mesabi range the rocks above the Virginia slate are not exposed. Thick glacial deposits conceal its relation to rocks and ores that are known in and near the Cuyuna Range. At the east end of the Mesabi, the later Keweenawan Duluth gabbro transgresses from the Virginia slate down across the iron-formation. Where the iron-formation emerges again, near Gunflint Lake and in a belt to Grand Portage, the Virginia slate (called Rove slate) is overlain almost conformably by Keweenawan sandstone and basalt flows (Grout and Schwartz, 1933).

Very much younger sediments occur at places along the Mesabi range from Coleraine to Eveleth (Grout, 1919, Fig. 27), unconformably overlying the slate and ore bodies. There are conglomerates and shales, the latter carrying Cretaceous fossils (Bergquist, 1944). Similar Cretaceous shales between the glacial drift and the slates have been found by drilling southwest of the Mesabi range and north of the Cuyuna district.

RELATION OF THE VIRGINIA SLATE TO THE BIWABIK IRON FORMATION: For many years it has been believed that the Virginia slate overlies the Biwabik formation conformably. Only during the last 25 years have some geologists doubted that the contact is conformable. Conditions at the contact are much more complicated than one could have suspected before 1910.

The greatest obstacle to a straightforward solution of this problem is the fact that there are no outcrops which show the contact of the two formations, except in the two open pits mentioned. The Virginia slate is so soft compared with most of the rocks underlying it that it does not crop out except at a very few places, as for example where the Duluth gabbro has metamorphosed the slate to a hornfels.

The upper few feet of Biwabik formation are also absent in known outcrops because they consist largely of easily weathered carbonates. Most of what is known about the contact is gathered by the inspection of diamond-drill cores. These are nearly all less than an inch thick. Anyone familiar with diamond drilling knows that much of the core, particularly if hard and soft rocks occur interstratified, is ground up and not recognizable by visual inspection. For this reason it has been impossible to find a piece of core which represents the actual contact of the two formations. The geologist must work, therefore, on indirect evidence. The problem is so important both to general geology and to ore finding that the facts are stated and discussed here in considerable detail.

Facts which bear on this problem and apply to the two formations as far west as Nashwauk (R. 22 W.) are presented as items 1 to 8. The data west of Nashwauk are given as items 9 to 16.

(1) The Biwabik formation in its uppermost 10–20 feet consists largely of carbonate rock low in iron but very high in calcite. In other words, a change in deposition from iron-rich taconite to a new kind of carbonate rock is definitely indicated as the Virginia slate is approached. The period of chemical precipitation of iron, which was long enough for the deposition of 500 and more feet of iron formation, was evidently approaching its end.

(2) The "Virginia slate," which is a hard shale and graywacke, is easily distinguished from taconite. It is dark gray and black and shows good bedding except in the graywacke portions. It is a clastic sediment and has a composition typical of shales and graywackes. It is high in alumina and low in iron (C. K. Leith, 1903, second analysis on p. 170). Its specific gravity is about 2.7 to 2.8.

(3) Where the Virginia slate is near or at the contact with the Biwabik formation, a little gritty and sandy sediment may be recognized in some drill cores. The thickness of this coarse-grained material does not exceed a few inches.

(4) There is *no* secondary oxidation or leaching of the iron formation at the contact with the Virginia slate unless there is also some decomposition (bleaching and kaolinization) of the slate.

(5) Where iron-ore bodies occur directly under the Virginia slate the slate may have been decomposed and bleached to a vertical distance of many feet. As much as 60 feet of altered slate has been observed in drill cores from rocks underlying fresh slate.

(6) The facts under items 4 and 5 indicate (if not prove) that the solutions which oxidized, decomposed, and leached the rocks circulated in them *after* deposition of the Virginia slate.

(7) As the Virginia slate is measurable in thousands of feet (one hole near Eveleth penetrates 1800 feet), its accumulation must have taken hundreds of thousands and perhaps millions of years.

(8) As circulation of *oxidising* ground waters most probably did not start until much of the Virginia slate had been eroded, another relatively long time interval elapsed before iron *ore* bodies, which were produced by deep oxidation and leaching, had their beginning. It is believed that this occurred during the Keweenawan period.

From Nashwauk west, conditions in part seem to contradict those toward the east.

(9) The granule-textured iron-formation, which is more than 600 feet thick east of Nashwauk, begins to thin westward and is about 400 feet thick not far from Calumet (R. 23 W.). It thickens from there westward and is 500 to 550 feet thick at Coleraine (R. 24 W.). For details *see* the longitudinal section (Gruner, 1946, Pl. I).

(10) The cherty and slaty taconites which form the upper part of the granule taconites are, at places, practically *all* oxidized and partly decomposed, forming "wash ores".

(11) This oxidation in many places reaches to depths of 200 to 300 feet in the iron-forma-

tion and in some drill holes extends nearly to the bottom of the Biwabik formation or to a total depth including drift and overlying slates of 900 feet.

(12) "Slates"—really fresh shales—overlie these oxidized taconites except where they have been bevelled by erosion at a relatively late date. These "slates" have been considered a part of the Virginia slate, but they are high in iron carbonate (Gruner, 1946, p. 50) and chemically are not Virginia slate but definitely a slaty division of the iron-formation.

(13) These iron-bearing "slates" which may be as thick as 200 feet gradually merge into the true Virginia slate above, from which they are distinguishable chiefly by the specific gravity of the drill cores or by chemical tests.

(14) The contact between these "slates" and the underlying oxidized granular taconites is not well represented by the available drill cores. The recovered pieces of core are, of course, the hard parts of the rock. The softer parts were ground up and lost or recovered as finely divided cuttings. Furthermore, 30 and more years ago when most of this drilling was done, drill runners and engineers were careless in their selection of pieces of core to keep, so that a 5-foot run of the bit may be represented by anything from half a foot to 2 feet of core.

(15) While the "slates" are "fresh" and unoxidized close to the contact with the cherty taconite, there is evidence that some small thickness of slate (2–4 feet at the most) is oxidized and decomposed. It is impossible to decide today without more careful drilling whether this oxidized slate, indicated by short, thin "discs" of drill core, is a part of the taconite proper or a part of the iron-bearing overlying slate. If it is taconite the oxidation and decomposition may be older than the deposition of the slates; if slate, the slates must have been deposited before oxidation, and would have guided the circulating solutions along the contact, not permitting them to penetrate far into the slates above.

(16) There is much evidence in parts of the Biwabik formation that slates guided the solutions. The Intermediate slate, for example, though only 10 to 30 feet thick, profoundly influenced the location of the oxidized and leached zones of the taconites and ores. The writer has seen many places where the lower part of the Intermediate slate is oxidized for a few inches to several feet. Above the oxidized portion, the Intermediate slate is unattacked. The taconite under the slate at such places is highly oxidized and altered for many feet vertically and long distances laterally.

With these facts before us, the question remains: What is the age of the oxidation of the taconite west of Nashwauk? If it is pre-Virginia slate then taconite was exposed near the surface, and no deposition occurred here while a thick layer of thin-bedded rocks was being deposited on the same formation a short distance to the east. This condition would have had to continue over a very long period of time, as oxidation is many hundreds of feet deep. The border line between deposition of Virginia slate and oxidation of iron-formation would have had to be quite stationary during all this time. Admitting for the present that such conditions could have existed, it taxes the imagination to believe it was followed by deposition of thick ferruginous shales over this oxidized material and a gradual transition to Virginia slate. The absence of any relatively coarse sediments on oxidized material and the sudden appearance of reducing conditions, and deposition of thick beds of iron carbonate with shales, when added to the other strange combinations make it very unlikely that an unconformity exists on the western Mesabi Range. It is very much simpler and more logical to assume that oxidation, decomposition, and leaching of the taconites happened at the same geologic time for the whole range. This must have been after deposition of the Virginia slate as demonstrated by items (1) to (8).

J. F. Wolff (1917, p. 165) was among the first to doubt the conformable relationship of the Biwabik formation and Virginia slate. He has expressed himself (personal communication) lately as being strongly inclined to think of an unconformity because of the evidence given in (9) and (10). He mentions some layers containing a few pebbles at or near the contact, but also points out that conglomeratic lenses may occur almost anywhere in the upper portions of the Biwabik formation where they certainly do not suggest a major unconformity.

We leave the discussion with the probability that the two formations are nearly conformable

in Minnesota. If there is an unconformity in the Minnesota Animikie it is more likely below the Biwabik than above; between the Pokegama and Biwabik, Leith and all recent students recognize a "minor erosion interval" but seem to agree it is a disconformity rather than an unconformity.

POSSIBLE IRON-BEARING BEDS IN THE VIRGINIA SLATE: Item 12 notes that there are certain beds in thick slates, rich in iron carbonate. This raises a question whether they are parts of a slate formation or of an iron-formation, a matter of great importance as a basis of prospecting for the iron-bearing beds. There is much difference of opinion as to whether there may be southwest of the Mesabi Range an increasing number of ferruginous beds in the Virginia slate. It has even been suggested that parts or all of the Cuyuna Range consists of such beds in the slate formation.

EXTENSIONS OF THE ANIMIKIE GROUP IN MINNESOTA: The correlation of the Mesabi range with the Gunflint range is satisfactory and is accepted by nearly all who are familiar with the ranges (C. K. Leith, 1903; Broderick 1920; Gill, 1926, p. 28–88). On the contrary the relation of the Mesabi to the Cuyuna range has aroused a great deal of discussion. Leith early sketched a probably connection, and the evidence is sufficient to convince some men but not others.

The points favoring correlation are (1) granule textures, known for many years and not known at any Lake Superior district in two formations of very different age; Thiel (1927) reported them from the Cuyuna, and they occur at several mines; (2) a sequence of (a) quartzite, (b) iron-formation with four members—lower cherty, lower slaty, upper cherty, and upper slaty (questioned by some)—and (c) an overlying slate, though the Cuyuna slates are commonly red and green instead of gray like the Virginia slate; (3) the rocks lie below the Keweenawan flows but have not been shown to be older than Algoman—evidence is meager on the Cuyuna; (4) the differences in rocks and ores can be explained as an effect of deposition of Cuyuna rocks near a shore line of sedimentation; (5) a bed that may have algal structure near the base of the lower cherty division is an especially good horizon marker. A good place to see the sequence of quartzite, Lower Cherty, Lower Slaty, is in the west end of the Maroco pit.

Against the correlation there are (1) the much greater metamorphism on the Cuyuna than on the Mesabi; the slates seem similar to those in the adjacent pre-Animikie Thomson slate; (2) the difficulty of connecting the magnetic belts of the Mesabi with the magnetic belts of the Cuyuna; (3) the very much greater proportion of true clastic sediments in the Cuyuna; (4) Zapffe (1926) objects to correlation because of evidence of an unconformity in some drill cuttings between the ranges, but these have been studied in detail and are probably Cretaceous; (5) the folding and deformation of the Cuyuna before the Keweenawan flows is much greater than the deformation of the Animikie on the Mesabi and farther east, suggesting that the Cuyuna may be an older series.

The tendency to use the petrographic character and sequence as a basis of correlation is strong, chiefly because granule textures and a sequence of cherty and slaty divisions have been long known in the Lake Superior Animikie, and no district has shown evidence opposed to correlation. Nevertheless we present the correlation as an old suggestion which is still interesting, but questionable.

A final suggestion that may deserve study is that the main thick iron-bearing series of the North range of the Cuyuna may be the equivalent of the Biwabik, whereas the long narrow belt in extensive slates of the South range may be an odd development in the Virginia slate. The data are too few to settle this problem.

Sioux Formation

The name of the Sioux quartzite was given by C. A. White (1870) to an extensive series of outcrops (in southwestern Minnesota and adjoining areas) of hard, pink, vitreous quartzite "assigned to the Huronian." Farther north in Minnesota the only similar rocks are those of the Pokegama quartzite. Correlation was tentative because outcrops are scattered and show little as to relations to other formations. The Cretaceous lies above the quartzite in sec. 27, T. 110 N., R. 30 W., along U. S. Highway 14.

Locally the Sioux quartzites grade into thin shaly beds, widely known and used as pipestone, and elsewhere into pebbly beds. The maximum thickness of the formation shown by wells is somewhat more than 500 feet (Thiel, 1944, p. 353, 401, 461). The structure, metamorphism, and most of the petrographic features suggest correlation with the Animikie, but some of the pebbles of the conglomerates (near New Ulm, near Courtland, and north of Luverne, in secs. 3, 11, and 14, T. 104 N., R. 45 W.) closely resemble the granule-textured iron-formation of the Biwabik. If the pebbly Sioux beds are younger than Biwabik they should not be correlated with the older Pokegama—the Sioux would be upper Animikie, or between Animikie and the Paleozoic. No such quartzite is known in the Virginia slate or Keweenawan, though the Barron quartzite of Wisconsin has been tentatively placed in the Keweenawan (Hotchkiss, 1915). The Sioux is, therefore, placed tentatively between the Animikie and the Keweenawan. It might be equivalent to the Sibley series (Tanton, 1931) which also has pebbles of Animikie iron formation.

Post-Animikie Unconformity

The relation of the Keweenawan to the Animikie is well studied in Wisconsin by Van Hise and Leith (1911, p. 234, 378) where broad field studies show that in the central part of the belt the Keweenawan rests on some thousands of feet of Tyler (Animikie) slate, but at each side it transgresses down through that slate and rests on the underlying iron-formation (Ironwood). The strikes and dips are almost parallel. Similar relations appear in the few exposures in Minnesota. A possible exception is in the Cuyuna district if the iron-bearing rocks are proved to be Animikie. The steeply dipping iron ores are said there (Van Hise and Leith, 1911, p. 215) to be overlain by surface volcanics.

Keweenawan Group

Nomenclature.—Since 1905 the spelling used by the International Committee (Hays *et al.* 1905), Keweenawan, has been generally adopted. Brooks (1876) first referred to it as Keweenawian. T. Sterry Hunt (1873, p. 339) used the spelling Keweenaw.

We prefer to consider Keweenawan and Animikie as two groups rather than to combine them into one large group (Pettijohn, 1937, p. 190). The Keweenawan rocks overlie the Animikie and underlie the Cambrian (in Minnesota the Upper Cambrian).

The lowest formation of the Keweenawan (or a formation below the Keweenawan and above the Animikie) in an area near Minnesota, Tanton (1927) called the "Sibley". At the type locality, Thunder Bay (Tanton, 1931, p. 47–50), the beds include about 250 feet of conglomerate, sandy and shaly sediments, and some limestones, overlying the Animikie with slight unconformity. This basal Keweenawan or older complex formation is not recognized in Minnesota. But *see* the Sioux quartzite, above.

Lower Keweenawan, Puckwunge formation.—The term Puckwunge conglomerate was first applied by N. H. Winchell (1897; 1899, p. 13, 297, 307, 325, 327, 517) to exposures in the valley of Puckwunge Creek (now Stump River) in eastern Cook County (sec. 25, T. 64 N., R. 3 E.). Later Winchell extended the term to cover the basal Keweenawan sediments from Puckwunge Creek eastward to Grand Portage Bay and also to somewhat similar sediments at the base of the Keweenawan cropping out principally along the St. Louis River just west of Duluth. Puckwunge formation is here considered better usage, because most of the exposures are sandstone.

Van Hise and Leith (1911, p. 367–394) used the term Puckwunge following Winchell but restricted it to the basal formation of the Keweenawan. The stratigraphic position of these sediments is not questioned. The formation rests unconformably upon Animikie and older slates and beneath Middle Keweenawan basaltic lava flows. The possibility that some Middle Keweenawan flows are earlier than basal Keweenawan sediments (as in Wisconsin) is not supported by studies of the accessory minerals in the Minnesota exposures.

DISTRIBUTION: The Puckwunge, and in fact the lower Keweenawan throughout the Lake Superior district, has a rather limited distribution. The exposures in Minnesota are in two

general areas. The most accessible are just west of Duluth from the valley of the St. Louis River northward about 5 miles to sec. 17, T. 49 N., R. 15 W.

The base of the Duluth gabbro is transgressive in a broad way, and from Duluth, where it intrudes flows, it extends north, bevelling the lower flows and then the Puckwunge (Schwartz, 1949). One hundred and thirty miles to the northeast in Cook County the basal Keweenawan formations gradually reappear. Outcrops of sandstone and occasionally conglomerate have been traced from sec. 22, T. 64 N., R. 3 E., where the gabbro transgresses into the underlying Rove slate, eastward in a narrow belt to Grand Portage Bay and the islands off Pigeon Point. (*See* Grout and Schwartz, 1933.) The easternmost exposure is on Lucille Island.

LITHOLOGY: The type exposures contain abundant quartz conglomerate and were described by Winchell (1899, p. 499).

In the belt ending at Lucille Island outcrops are reasonably continuous along a bluff but are almost wholly of the sandstone facies. The base is usually concealed in a swamp north of the bluff. The sandstone is nearly white, uniform, and only moderately cemented. Some grains are well rounded, but many are angular. Cross-bedding is common (Grout and Schwartz, 1933, Fig. 22). A few inches beneath the overlying basalt flow the sand is almost quartzitic; otherwise differences are slight. A 50-foot section in sec. 1, T. 63 N., R. 5 E., contains fine-grained, thinly bedded gray to buff sandstone.

Feet	
0–15	—a fine-grained, generally massive, gray sandstone with some thinly bedded and some shaly layers. No baking effects.
15–28	—a fine-grained, thinly bedded, gray, slightly shaly sandstone.
28–38	—a fine-grained, bedded, buff sandstone.
38–50	—a fine-grained, gray, somewhat shaly sandstone.
50	—talus—bottom of hill.

The largest exposures of Puckwunge in the Duluth area form a west-facing bluff in secs. 17 and 20, T. 49 N., R. 15 W. Above a few feet of white quartz-pebble conglomerate is buff sandstone, cross-bedded, and with grains fairly well rounded. Within a foot of the overlying lava flow the sandstone is silicified and at places is banded dark and light. The main portion of the sandstone strongly resembles that in the Grand Portage region.

THICKNESS: The Puckwunge formation represents the whole of the Lower Keweenawan as defined by Van Hise and Leith (1911). Thickness probably does not exceed 200 feet and at places is much less.

STRUCTURE: The structure of the Puckwunge formation is simple throughout. It lies on the north limb of the Lake Superior syncline and strikes generally east-west in the Grand Portage area. Near Duluth the strike is nearly north in sec. 20, T. 49 N., R. 16 W., in conformity with the regional structure. The dips are usually gentle, at places less than 10°, and range to an observed maximum of 17°.

RELATIONS TO OTHER FORMATIONS: Contacts are very poorly exposed, but the sandstone facies is well shown at places not far from exposures of the underlying formations.

As pointed out in the section on the Thomson formation, the relation between the two and comparison with the Rove formation are matters of considerable significance.

Throughout its extent in Cook County the exposed Puckwunge is mainly a sandstone dipping 12°–16° S. with no visible discordance from the Rove formation below. The "platy" beds of the Rove lack secondary cleavage but are more indurated than the Puckwunge sandstone above. A white sandstone could hardly be derived from impure graywacke. The induration and the abrupt change to a conglomerate containing a considerable assortment of pebbles make it reasonably certain that an erosional unconformity separates them.

In the area west of Duluth the Puckwunge rests on the Thomson formation with marked angular unconformity. Moreover, the Thomson has a highly developed slaty cleavage, whereas the Puckwunge is not metamorphosed. The dips of the Thomson at or near the overlying conglomerate range mainly between 45° and 90°, whereas the dip of the overlying sandstone does not exceed 10°. The most instructive exposure is along the bluff in the NE.¼NW.¼ sec. 20, T. 49 N., R. 15 W.

The exposures in Cook County are well shown on the detailed maps of Minnesota Geological Survey Bulletin 24. Many of these are

relatively inaccessible, but the important exposures on Grand Portage and Lucille islands may be visited by motor boat from Grand Portage Village; and the sandstone and overlying flows are exposed in sec. 8, T. 63 N., R. 6 E., about a mile southwest of Grand Portage Village.

Middle Keweenawan basic and acidic extrusives: Keweenawan Point volcanics.—The Keweenawan lava flows have usually been referred to by that term using the general time unit for the formation. Grout and Schwartz (1939, p. 13) suggested that Keweenaw Point volcanics seemed the logical formational name for the flows and associated fragmental rocks of the middle division of the Keweenawan.

Extrusives, mainly lava flows, of the Middle Keweenawan underlie a very large area in northeastern Minnesota extending from the most southerly known occurrence in a drill hole at Stillwater to the islands which lie south of Pigeon Point in Lake Superior. The distance between these points is 260 miles in a straight line which passes through a portion of western Wisconsin.

The flows are probably continuous at depth throughout the entire length, but they are widely covered at the south end by Cambrian beds, and farther north by Upper Keweenawan sediments and generally by widespread deposits of glacial drift. There are several faults which offset the flows but may not break the continuity.

The occurrence at Stillwater is known from a single deep drill hole where a series of flows was penetrated from a depth of 3175 feet to the bottom of the hole at 3500 feet. In Minnesota and Wisconsin, flows are not known east, south, or west of this point.

The enormous volume of the flows suggests a deep source such as a basalt zone below the crust. Hotchkiss (1923, p. 672) thinks they rose in the bottom of the Lake Superior syncline and may have been fed from a batholith, though no such mass has been exposed, and most batholiths are not of basaltic composition.

Tanton (1931, p. 64, 86–87) believes that the flows of the north shore near Fort William and Thunder Bay came from the south, perhaps from some vent near the center of the present Lake Superior basin. Along the Minnesota coast, the bent amygdules give conflicting evidences, but perhaps the vents may have been largely south of the flow outcrops.

RELATION TO OTHER FORMATIONS: The general situation at the base of the Keweenawan has been discussed in some detail under the heading of Puckwunge formation and need not be repeated here. Near Duluth and Grand Portage the lowest exposed flow lies on the Puckwunge sandstone. Elsewhere the base of the flows is not exposed. Over most of the Keweenawan area of Minnesota glacial drift is the only younger formation exposed. In the deep well at Stillwater 2458 feet of Fond du Lac sandstones occurs above the flows and below the beds definitely Cambrian in age.

At and near Taylor's Falls on both sides of the St. Croix River basalt lava flows are directly overlain by fossiliferous Cambrian beds of the Dresbach formation. Farther north in southern Pine County also the Dresbach formation lies on the flows, but there are only small exposures. The westerly limit of the flows in the area south of Duluth is a large fault which was studied by Frey (1939) and Welsh (1941) in recent years by geophysical means. Flows are reported under the drift, lying on the Cuyuna iron-bearing rocks near Brainerd.

Near Duluth the flows are above the gabbro intrusive, except for a few in sec. 20, T. 49 N., R. 15 W., previously described. The upper portion of the group is beneath or south of Lake Superior. It is well known on Keweenaw Point, Michigan, and in northern Wisconsin, where the base is a fault, but the gradation into Upper Keweenawan sediments is widely exposed.

STRUCTURE: The structure of the Keweenaw Point extrusives in Minnesota is normally rather simple. The area northwest of Lake Superior is the north limb of the great syncline and has at most places a dip of 10°–15°. The principal departures from this gentle dip are a result of the intrusives described below—a local reversal of dip may be seen northeast of Grand Marais. Recognizable faulting is not common. Several maps indicate a fault in the lake near the north shore, but direct evidence is lacking, and the evidence cited for it may be as well explained by the gentle dip of the flows and sills toward the lake.

LITHOLOGY: The lava flows of the Lake Superior area have long received intensive petrographic study, and it is unnecessary here to describe them in detail. N. H. Winchell (1900) described thin sections from all along the Minnesota coast. More recently their character has been summarized by Grout and Schwartz (1939) and by Schwartz (1949). The flows of the type area on Keweenaw Point, Michigan, are described in considerable detail by Butler and Burbank (1929).

While the flows in Minnesota vary in detail they may all be classified in a general way as basalt or felsite (rhyolite). The basalts greatly predominate, but at intervals along the coast there are considerable areas of felsites (some of them porphyritic), perhaps 10 per cent of the total thickness of flows. It seems characteristic to have considerable groups of each rather than an alternation of single flows. Both basalts and felsites are well exposed along the shore within the corporate limits of Duluth. Flows of an intermediate composition are rare or lacking. There is variety in the proportion of light and dark minerals, but a scarcity of intermediate feldspars, such as would warrant the use of the term andesite as now defined by most petrographers. There is considerable variety in textures, and Lane's (1911) terms for types apply about as well as they do in Michigan.

The abundant basalts in Minnesota are ophite, melaphyre, porphyrite, and glomeroporphyrite. For alternation of these rocks in the sequence the reader is referred to Sandberg's tabulation (1938).

The total thickness of flows on the north limb of the Superior syncline will always be unknown because many of the later flows are hidden beneath the lake. Sandberg's very careful measurements from the base of the Middle Keweenawan at Duluth for 20 miles northeast gives an estimate of extrusives and related sediments of 20,856 feet. Grogan has continued the sequence for 16 miles more and gives an estimate of 3207 feet omitting intrusives.

From Split Rock (Sec. 5, T. 54 N., R. 8 W.) to Two Islands 5 miles east of the Lake County-Cook County line (sec. 10, T. 58 N., R. 5 W.) the strike is, in general, only at a slight angle to the shore, but the distance is about 40 miles. Perhaps an estimate of 30,000 feet would be of the right order for the total thickness of the flows on the north shore of Lake Superior with an unknown thickness under the lake.

Middle Keweenawan basic and acidic intrusives: Duluth gabbro.—Intrusive action in Middle Keweenawan time was on a grand scale, and the major mass, known as the Duluth gabbro, is one of the largest known floored intrusives. Besides this there are a great number of sills and dikes, large and small. Those in the series of flows are known as the Beaver Bay complex, and those in the underlying Rove (Animikie) slates are called Logan intrusives (but dikes pass from one area to the other suggesting that they are of nearly the same age). The late granites associated with the basic rocks may be of the same age as certain batholiths of central and southwestern Minnesota.

The Duluth gabbro was early studied and named by Irving (1883, p. 124–134). At Beaver Bay (T. 55 N., R. 8 W.) a large sill intrudes the series of Keweenawan flows, but there has not always been a recognition of the dual nature of the exposures. Irving (1883, p. 143–146) gave the name Beaver Bay Group to all the rocks including flows, sills, dikes, and others. Elftman (1898, p. 183) and Winchell used the same expression, with about the same meaning. Since the flows are part of a single series, the need for a special term concerns only the intrusives, and Grout and Schwartz (1939, p. 13) called them the Beaver Bay Complex. The Logan sills were named by Lawson (1893, p. 48) and defined as intrusives into the Rove (Animikie) slates. The sills grade into dikes, and the whole group may well be called Logan intrusives.

The Duluth gabbro formation has been studied by several geologists, and a summary of its petrography was given by Grout (1918b). It had earlier been referred to as a sill, laccolith, or flow, but its structure and probable form differ from these, and Grout (1918a, p. 518) made the Duluth gabbro the type of a lopolith. Much of the mass is in layers of slightly differing mineral content, but the mass is very large, and early structures have been confused by late movements and injections, so that fragments, large and small, are included and at places partly dissolved in later magma.

PETROGRAPHY: Nearly all the Middle Ke-

weenawan intrusives near Lake Superior have an average composition that is basaltic (diabasic or gabbroic), but the large masses are notably differentiated. Small dikes are fine-grained basalts and may have glassy borders. Many dikes and sills from 10 to 300 feet thick show diabasic textures with grains about 1 mm. in diameter (dolerites), and a good many have coarser plagioclase phenocrysts. Phenocrysts tend to concentrate near the tops of the larger sills (as at the Leveaux Mt. Lookout (T. 59 N., R. 4 W.) between Tofte and Lutsen). An easily accessible sill with such phenocrysts crops out on the north shore of Loon Lake (T. 65 N., R. 3 W); there are associated some feldspar aggregates or fragments of anorthosite. The larger sills more than 500 feet thick show a great variety of differentiated facies.

At Pigeon Point, the northeast corner of the State, the sill is olivine diabase near the base, quartz diabase near the top (exactly at the tip of the point the rock shows early olivine and later micrographic intergrowths of quartz and orthoclase).

The Beaver Bay sill has a similar series of facies. There are, on the shore of Beaver Bay (T. 55 N., R. 8 W.), an unusual number of inclusions of fragments large and small in the diabase. The dominant inclusions are anorthosite (Grout and Schwartz, 1939, p. 48–49) and a red rock that seems to be partly rhyolite but may have been much modified if not really melted by the heat of the basic magma.

Three sills at Duluth were studied in detail by Schwartz and Sandberg (1940, Table 1) with analyses.

The Duluth gabbro shows the greatest petrographic variety (Grout, 1918b). There is a great volume of olivine gabbro and normal gabbro in layers from an inch to many feet thick, and there are scattered layers of other compositions.

In the more extensive but less accessible areas of gabbro in the belt northeast of Duluth there are similar facies, and the red rock makes up a much larger percentage of the total intrusive. It resists erosion even better than the gabbro and is the rock of some of the highest peaks in the State—Brule Mountain and Eagle Mountain, in Cook County (T. 63 N., Rs. 1 and 2 W.).

One sill-like granite mass, with small apophyses in the gabbro, extends from Jack Lake on Temperance River west to Smoke Lake in Cook County. It has a texture more like a pink aplite than like the granophyric red rock. A few miles farther west near Dent Lake a still different granitic mass grades through granodiorite to gabbro and still seems to be different from the common red rock.

The gabbro, in addition to the normal granitoid rock, has local pegmatitic segregations (Grout, 1918c) and dikes near the base.

Most of the intrusives are fairly fresh, but there are locally some evidences of hydrothermal mineralization—nickel and copper prospects are widely scattered, but none have proved valuable. Weathering commonly turns the olivine-brown and the gray feldspar white.

Succession: The intrusives must have cut underlying formations to feed the first of the flows; and, from that beginning, probably continued intermittently to the end of the last flow. No intrusives are known to cut the sediments above the lavas, either in Minnesota or in the wide exposures in Wisconsin (Tyler *et al.*, 1940, p. 1465) though Irving says (1883, p. 441) one or two dikes are late enough to cut some upper Keweenawan sandstones. Nearly all the large intrusives have been cut by later intrusives dikes, and it is certain that some early dikes were cut by the large intrusives. The sequence of intrusion is thus complicated but seems to have been completed in Middle Keweenawan time. The following are selected as important or easily accessible exposures that indicate successive intrusions.

East of the railway tunnel near Short Line Park at Duluth the gabbro sends dikes and stringers into the Keweenawan lavas of its floor; and east of Duluth Heights the top of the gabbro sends stringers into its roof of similar flows.

Trap dikes cut the upper part of the gabbro along the bluff in the center of the city, just west of City Park, and are so late that they have chilled margins, indicating that the gabbro was not only solid but cooled. Similar dikes cut the gabbro at Little Saganaga Lake in Cook County (T. 64 N., R. 5 W.). Less accessible but very indicative that some trap dikes are late, are small dikes that cut the

granite and granodiorite which cut the gabbro in west-central Cook County; they can be seen along the water's edge in the outlet bay of Zenith Lake, T. 63 N., R. 5 W.

At several places Logan sills intruding the Animikie sediments are recrystallized by the heat of later gabbro intrusion so that the diabase is changed to a sugary hornfels. Since other formations are also changed to hornfelses with similar appearance, this sort of evidence might not ordinarily be convincing, but two places prove that some of the sills antedated the gabbro. There is a sill in the iron-formation near Dunka River on the Eastern Mesabi range (secs. 26 and 35, T. 61 N., R 12 W.) which has not only the hornfels texture near the gabbro, but also some characteristic phenocrysts of plagioclase, so that there can be no doubt it was a Logan intrusive. Also there are Logan intrusives south of the east end of Mayhew Lake (along the Gunflint Trail), secs. 33 and 34, T. 65 N., R. 2 W., which can be traced almost continuously from a hornfels at the gabbro contact, along a ridge to a belt of diabase without recrystallization, half a mile from the gabbro (Grout and Schwartz, 1933, p. 8, Pl. 2).

Presumably the intrusives came from the same deep source as the great flows of the Keweenawan. Possibly there were subsidiary reservoirs, and it has been suggested that some flows were derived from the Duluth gabbro magma, which must have constituted a great reservoir. There is little evidence.

The gabbro itself seems structurally to be 3 miles thick at Duluth and possibly 10 miles thick in its widest part in Lake County. Such a mass can hardly be explained as an aggressive intrusion—rather it must have developed by a sinking of the floor, into a still larger volume of magma below.

The gabbro intrudes the Keweenawan flows at Duluth and at the northeast where it fingers out. Between them, along the floor of the major body of magma, it overlies in succession the Animikie (on the East Mesabi and Gunflint), the Algoman granite of the Giants range, the Knife Lake group (near Knife Lake), and the Keewatin greenstone (at Disappointment Mountain and Gabimichigami Lake). This might be taken to mean a transgressive intrusion, but another interpretation is more probable. The intrusion is lens-shaped, and both ends spread near the bottom of the Keweenawan. Probably erosion in post-Algoman time, and possibly some after Animikie time, left this series of formations exposed for deposition of the basal Keweenawan. If so the later gabbro-intrusive had only to transgress across a few flows to reach the unconformity, where all the rocks would be in contact with the gabbro.

THICKNESS: The intrusives range from a fraction of an inch to several miles in thickness. The Duluth gabbro is slightly faulted, but there is no recorded evidence of a repetition of layers. If the dip of its structure and the width of its outcrop area are used to calculate its probable thickness, it is more than 10 miles thick in Lake County and more than 3 miles thick near Duluth. Associated and probably related sills add roughly another mile.

DISTRIBUTION AND RELATIONS: The Duluth mass extends in a great crescent from the western part of the city of Duluth, north and northeast through its widest exposure in Lake County, to its Cook County limits where it divides into sill-like fingers; one finger reaches Lake Superior near Hovland (T. 62 N., R. 4 E.), and another ends near Portage Brook (on the west side of sec. 27, T. 64 N., R 3 E.); the southern finger for the last few miles shows little but gabbro, but the northern one is dominantly red rock with a little gabbro along its north side.

The next, smaller intrusives extend, as the Beaver Bay complex, from Duluth northeast to Grand Portage cutting the area of Keweenawan flows at many places. Northeast of this (as the Logan intrusives) they cut the Animikie rocks along the boundary belt from Pigeon Point to Gunflint Lake, and in outliers in the Mesabi range.

The still smaller dikes and scattered intrusives of irregular and uncertain forms extend over these areas and even more widely. Fresh trap dikes are known from Stearns County north to Lake of the Woods, and from that area east. They thus range over the northeast quarter of the State and probably beyond. They have commonly been assumed to be Middle Keweenawan because none are known to cut Upper Keweenawan or later sediments, and such dikes do cut all formations older than

those sediments. They are not all in one regular uniform system, but are influenced by the joints in the earlier rocks.

The acidic intrusives of the Keweenawan are not exclusively the "red rocks" associated as upper facies of gabbro and diabase. From the Canadian boundary to the Minnesota River Valley there are scattered granites assigned for several reasons to the Keweenawan. There are in the first place accessory zircons of a character not known in any rocks older than Keweenawan—the "normal" zircons of simple crystal form, little zoning, and ranging from colorless to pale brown (Tyler *et al.*, 1940, p. 1465). In the next place they are cut by trap dikes as most of the red rocks are. Furthermore, several show intrusive relations to older intrusives.

The red St. Cloud granite and related Cold Springs and Rockville granites (T. 123 N., R. 29 W.) and Crystal Gray quartz monzonite, which furnish large amounts of ornamental and monument stone, are believed to be Keweenawan.

CONTACT EFFECTS: The large intrusives produce contact metamorphism of considerable extent (Grout, 1933a; Schwartz, 1924; 1943). The gabbro makes a hornfels zone some hundreds of feet wide from several of the formations in its floor—notably the Keweenawan basalts, the Animikie slates, the Knife Lake slates, and the Keewatin greenstones. It is hardly possible to identify the original formation of one of these hornfelses except by tracing it to an unaltered exposure. Smaller dikes and sills make local hornfelses; a noteworthy occurrence is that between two large diabase dikes cutting the Rove slates (Grout and Schwartz, 1933, p. 27).

Correlation of Keweenawan intrusives is fairly satisfactory. The main body of gabbro, though concealed by drift in some places for miles, has outcrops enough and an internal structural trend so clear that there is little doubt of its continuity from Duluth to Portage Brook, and to Hovland on the shore of Lake Superior in Cook County.

It is a reasonable inference also that it connects below Lake Superior with the gabbro in Wisconsin.

Excellent exposures for study of the Keweenawan intrusives are (1) at Duluth (Grout, 1933b, p. 69) showing banding of a variety of facies, the relations of red rock, and the hornfelses produced by contact action on underlying flows; (2) along the highway near Beaver Bay (Grout and Schwartz, 1939) showing a large sill with irregular contacts, dip slopes, and inclusions of anorthosite and red rock; (3) near Grand Portage showing ridges resulting from dikes cutting the Rove formation; (4) at Carlton (T. 48 N., R. 17 W.) showing similar dikes cutting the Thomson slate; and (5) at St. Cloud, Stearns County, where granite from the quarries shows trap dikes in considerable abundance, and the red granites are themselves believed to be Keweenawan. Less accessible but significant are the dikes cutting the island rocks on Rainy Lake and Lake of the Woods; the intrusive granite and granodiorite in west-central Cook County; and the East Mesabi and Cuyuna areas where dikes and sills cut the iron-formation.

Fresh trap dikes cutting older formations through the Lake Superior region are commonly referred to Keweenawan age, but the ages of several may be questioned. The dikes cut nearly all older formations. On Rainy Lake (T. 71 N., R. 23 W.) prominent dikes cut through the Keewatin greenstone and Knife Lake (Seine) sediments about 5 miles east of Ranier. On the west side of Cache Bay of Saganaga Lake, not far from the Minnesota line in T. 66 N., R. 6 W., a black dike cuts the pre-Knife Lake granite and can be followed into the Knife Lake sediments. Near Snowbank Lake (T. 63 N., R. 9 W.) trap dikes cut both the Knife Lake group and the granite stock which appears to be Algoman. Gruner (1941, Pl. 1) shows diabase dikes cutting the Knife Lake group around the east bay of Ensign Lake, and several other places. Trap dikes cut the Algoman Giants Range granite along the road north of Virginia (T. 59 N., R. 18 W.). Mention has been made above of the sills and dikes of Keweenawan diabase in the Animikie (Mesabi and Rove series). The East Mesabi taconite (T. 60 N., R. 13 W.) of the Biwabik formation is cut by several trap dikes. Trap dikes cut even the Keweenawan Duluth gabbro; some may be seen at City Park in Duluth; and some as far east as Tuscarora Lake in Cook County.

Upper Keweenawan.—The upper Keweenawan group is divided into a lower part called Fond du Lac beds and an upper part called Hinckley sandstone. Both these terms were first used for rocks of geographical occurrences in Minnesota without any very definite evidence as to their relations to Cambrian or Precambrian known rocks. Both are older than the Dresbach formation as was stated in the original usage of the terms. Their relations and contacts are well determined by the data in the following pages.

Leith, Lund, and Leith (1935, Pl. I) place the Duluth gabbro and its upper "red rock" granitic facies (and the Killarney granite) *above* the Upper Keweenawan sediments, but mark them "exact position in the scale doubtful". Outcrops near Duluth make it fairly clear that the Upper Keweenawan sediments as here described are later than the gabbro and its red-rock facies. The intrusive lopolith as a whole transgresses only a little across the Middle Keweenawan flows; it was intruded following the general structure of the flow series, depressing them into a pronounced syncline—it was too thick and heavy to have stood up as a dome. If this intrusive had come later than the deposition of Upper Keweenawan sediments, they would have been similarly deformed. The outcrops from Fond du Lac to Hinckley show no such prominent syncline.

FOND DU LAC BEDS: The term Fond du Lac was used by Upham (1884; 1899) for the dark-red to pink arenaceous shales and red to brown argillaceous to arkosic sandstones which form conspicuous cliffs along the St. Louis River southwest of Duluth. He defined them as above the Thomson slates. He applied the same term to the outcrops in the banks of the Snake River north of Mora in Kanabec County. The name Red Clastic beds or series was used by Hall, Meinzer, and Fuller (1911). They regarded it as a convenient term to cover those red beds of uncertain age penetrated in the deeper wells of southern Minnesota rather than as a permanent name for a series or a formation. There is now little doubt, however, that the red beds referred to as Red Clastics are equivalent to the red beds of the Lake Superior district which Winchell referred to as the Fond du Lac sandstone. Since these beds are typically exposed along the St. Louis River and in the quarries at Fond du Lac, the name Fond du Lac sandstone seems most appropriate. Upham (1901) specified that these beds were below the Hinckley sandstone. Wilson (1939b, p. 243) did not find any Upper Keweenawan in the Northwest Subprovince of the Lake Superior region, but we claim good evidence that these sediments overlie and contain fragments of the flows.

The Fond du Lac beds are composed of interbedded massive sandstones and thinly laminated shales. Some of the sandstone strata are slabby and thin-bedded with seams of bluish-green and red shale. Pebble horizons are common, and locally there are very minor amounts of clear, white sandstone. The massive sandstone strata range from 1 to 8 feet thick, and the shale facies are generally less than 5 feet thick. Locally, however, they are as much as 25 feet thick. The formation as a whole has a thickness ranging from several hundred to more than 2000 feet.

The sandstones are conspicuously cross-bedded, and some ripple marks are well preserved. The associated red and mottled shales have mud-crack patterns and a few indefinite trails that may or may not be of organic origin. The strata are only slightly disturbed from their original position and are scarcely affected by metamorphism. Some of the mica along the parting planes in the slabby sandstone may be a product of anamorphism. The beds dip from 3° to 10° SE.

DISTRIBUTION: The geographic extent of the formation is not definitely known. An outcrop has been reported as far north as Floodwood in St. Louis County, but this observation has not been verified. Southward from the outcrops near Duluth, on the St. Louis and Snake rivers, the formation has been penetrated by a number of deep wells which indicate that it is continuous in the Precambrian trough that extends generally north-south across east-central and southeastern Minnesota. It extends westward beyond Mankato (R. 26 W.) and eastward beyond Rochester (R. 14 W.) where it is 2033 feet thick. East of Rochester it thins rapidly, and at Winona (R. 7 W.) a number of deep wells pass from the St. Croixian sediments directly into Precambrian granite. Its greatest recorded thickness in Minnesota is at Stillwater

where a deep well penetrated 2458 feet of red sandstones and shales beneath the Cambrian and Hinckley sandstones.

For a section about 230 feet thick in the bluff of the St. Louis River in the SE.¼ sec. 6, T. 48 N., R. 15 W., near Fond du Lac, Minnesota (diagram in Tyler *et al.*, 1940, p. 1513–1515), *see* Minn Geol. Survey Bulletin 32, page 16.

Near Mora the Fond du Lac beds consist of coarse gray to yellow and red sandstone, together with conglomerates and argillaceous red shales. The sandstones form a low cliff along the east bank of the Snake River, about a mile north of the Mora cemetery gate. Most of the outcrops of shale occur still farther north where the shales were excavated in test pits 50 to 75 feet east of the river bank.

CONTACTS AND RELATIONS: The geographic relations along the St. Louis River suggest that faulting has displaced most of the basal beds of the Fond du Lac sandstone to a position far below the level of the outcrops of the Middle and Lower Keweenawan strata in that region. These relations are shown on the map in Figure 9. The Thomson slate is exposed in the bed of Little River just north of the center of sec. 1, T. 48 N., R. 16 W., and other exposures of slate continue at intervals up Little River as far as the tracks of the Northern Pacific Railway in sec. 35, T. 49 N., R. 16 W. Downstream from the slate exposure, at the mouth of a small creek which joins Little River from the north, is an exposure of quartz-pebble conglomerate. The exact contact between slate and conglomerate is not exposed here, but a similar conglomerate lies directly on the slate in the bed of the St. Louis River in sec. 15, T. 48 N., R. 16 W. (Fig. 9). The accessory minerals in the matrix of the conglomerates at the two locations are very similar, and both are undoubtedly Lower Keweenawan. A short distance downstream from the conglomerate on Little Creek in section 1 is an outcrop of red, arkosic conglomerate and sandstone that has the lithologic and petrographic characteristics of the Fond du Lac beds, including some basalt pebbles. Quite possibly a fault passes between the two outcrops and the Upper Keweenawan is in contact with the Lower Keweenawan along the fault plane. This interpretation is supported by the results of the drilling for the foundation of the dam on the St. Louis River in SW¼ SW¼ sec. 6, T. 48 N., R. 15 W. A drill hole penetrated alternating red shale and sandstone beds from elevation 606 feet to 504 feet without reaching the Lower Keweenawan conglomerate or the Thomson slates. The dam site is not more than three fourths of a mile from the conglomerate in section 1, but the unconformable surface drops an unknown amount more than 180 feet in that short distance.

Both Winchell (1899, p. 13) and Thwaites (1912, p. 70, Pl. XV) reported faults across the valley of the St. Louis River. Thwaites included a photograph of a normal fault with a displacement of 14 feet in the arkosic grits and sandstones in the gully just above the railway bridge in sec. 1, T. 48 N., R. 16 W. He states that another normal fault of at least 40 feet displacement cuts the red shales and sandstones exposed in the north bank of the St. Louis River near the line between St. Louis and Carlton counties.

Quite possibly the grayish-green sandstone with pebbles of Middle Keweenawan basalt that crops out along Little Creek near the quartz-pebble conglomerate in section 1 is equivalent in age to the interflow sandstones that occur farther north in the region of the Short Line Park. If so, then the Lower, Middle, and Upper Keweenawan sandstones are in direct contact with each other in secs. 1 and 6, T. 48 N., R. 16 W. Possibly the Middle Keweenawan sediments may lie disconformably over those of the Lower Keweenawan, and the Fond du Lac beds of the Upper Keweenawan are faulted down in justaposition with the older strata. Since no basaltic flows are known to extend southwestward beyond sec. 32, T. 49 N., R. 16 W., some sediments of Middle Keweenawan age may have been deposited here around the outer margin of the flows.

It is equally difficult to draw a sharp boundary or contact at the top of the Fond du Lac beds. The characteristic red color of these clastic sediments is not in itself a safe criterion for establishing the lithologic boundary between them and the overlying Hinckley sandstone. However, a detailed study of numerous drill cuttings and outcrop samples from widely separated areas has shown a marked difference in

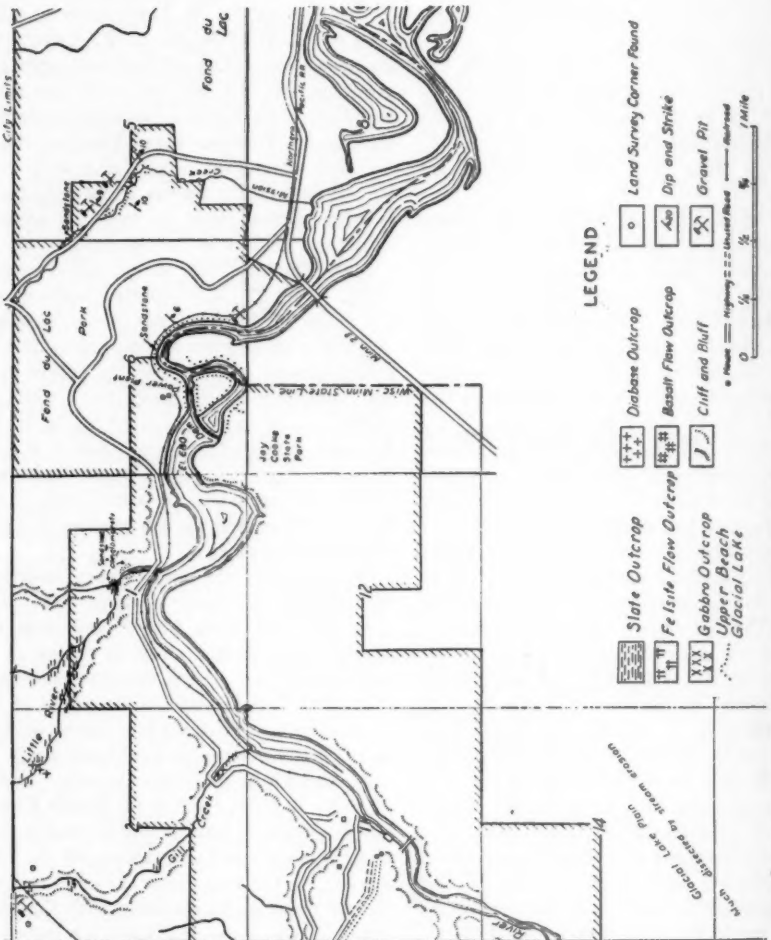

FIGURE 9.—KEWEENAWAN SEDIMENTS ON OLDER ROCKS NEAR DULUTH

the quantity and character of the feldspar in the two formations (Crowley and Thiel, 1940). These differences together with those of the heavy-mineral assemblages (Tyler *et al.*, 1940, p. 1508) are so uniform that they may be used for correlative purposes as well as for delimiting the red, clastic, Fond du Lac beds as a formational unit.

HINCKLEY SANDSTONE: The Hinckley sandstone was described by Winchell (1884), but the name was not proposed until later (1886, p. 337; 1888, p. 286). The type section of the formation at Hinckley (T. 41 N., R. 21 W.) exposed a ledge about 9 feet high which extended for 250 feet or more along the railroad just north of the Grindstone River. The sandstone is coarse to fine, usually medium-grained, salmon-pink to yellowish, and red, or nearly white. The colors are due to varying amounts of iron, which may form part of the cementing material and which occurs as either limonite or hematite. At many places the rock is well indurated and hence suitable for building stone, curbing, paving blocks, and even crushed stone. The fact that the formation contains very little feldspar and very few accessory minerals suggests that it was derived from an older sandstone that had previously passed through several cycles of sedimentation.

DISTRIBUTION: The Hinckley sandstone crops out almost continuously for nearly 20 miles along the Kettle River in the southwestern part of Pine County. Near Sandstone it forms castellated bluffs 50–100 feet in height above the valley. Some of the tributaries of the Kettle River have cascades or waterfalls which locally tumble 50 feet or more over the resistant beds of the sandstone. Beyond the major valleys the rock passes beneath the glacial drift which is not cut through by the smaller streams.

In the quarry wall at Sandstone, where a section of 100 feet of formation is exposed the strata range from a few inches to many feet thick. The lower half of this section contains thin beds of shale, some of which have been broken and recemented to form intraformational conglomerates. In the old quarry at Hinckley the beds are approximately horizontal, but farther northeast the strata dip from 3° to 5° SE. Cross-bedding is common but not conspicuous. In the vicinity of the upper falls north of Sandstone, the cross-bedding is inclined southward at 10°–30°.

In Pine County the Hinckley sandstone is more than 500 feet thick. The cliffs along Kettle River expose 100 feet, and the well at the Federal Prison south of the city penetrates 465 feet of sandstone which must be assigned to this formation. Southward from its type locality, however, the Hinckley thins progressively; it averages only about 150 feet near Minneapolis and St. Paul.

South of the outcrops in Pine and Kanabec counties, the Hinckley formation passes beneath the Cambrian sandstones of the St. Croixian series. Cambrian strata are encountered above the Hinckley in the wells of southern Chisago County, and the deep wells as far south as Faribault, Rice County, Mankato, Blue Earth County, and Rochester, Olmsted County, encounter a sandstone similar to or identical with the Hinckley, which occurs between the red Fond du Lac beds and the lowest sandstone (Mt. Simon) of the St. Croixian series. East of Rochester the Hinckley sandstone thins rapidly and pinches out against the westward-dipping surface of the Precambrian granite which occurs at a depth of approximately 500 feet at the city of Winona.

Good exposures of the Hinckley other than those near Hinckley and Sandstone may be observed along the Net River north of Holyoke, along the Willow River (sec. 32, T. 45 N., R. 19 W.) several miles east of the village of Willow River, and along the Kettle River near Rutledge (sec. 3, T. 43 N., R. 20 W.). An excellent artificial exposure that is very accessible occurs along the highway cut west of Askov where the road descends into the valley of the Kettle River (sec. 26, T. 43 N., R. 20 W.).

The contact between the Hinckley sandstone and the Fond du Lac beds is not exposed; hence it is known only from well samples and well drillers' logs. Apparently the pink and red color of the Hinckley grades downward into the Fond du Lac red beds without very great or sudden change in megascopic character. The well at the Federal Prison near Sandstone illustrates this point. Here the Hinckley sandstone is red for at least 350 feet above the base.

In general the physical and optical characteristics of the accessory minerals in the Hinckley sandstone are very similar to those of the Fond

FIGURE 9.—KEWEENAWAN SEDIMENTS ON OLDER ROCKS NEAR DULUTH

du Lac beds. Outstanding differences exist, however, in the types of zircon grains and in the percentage distribution of the various minerals. The order of their abundance is as follows:

Fond du Lac beds	*Hinckley sandstone*
1. Leucoxene aggregates	Zircon
2. Apatite	Leucoxene
3. Tourmaline	Tourmaline
4. Zircon	Opaque ores
5. Opaque ores	Garnet (very little)
6. Garnet	(No apatite observed)

The most conspicuous differences are that apatite is absent and garnet very rare in the Hinckley sandstone. Leucoxene is common throughout the Hinckley, but the floods of earthy aggregates that characterize the heavy fractions of the Fond du Lac beds were not observed in any of the Hinckley samples.

The top of the Hinckley sandstone is not known to be exposed in Pine County. In the area of outcrops along the St. Croix valley north and south of Taylors Falls and east of Pine City the basal members of the St. Croixian series, the Mt. Simon and the fossiliferous Eau Claire, rest unconformably on the lava flows of the Middle Keweenawan.

Drill cuttings from wells west of the valley in Pine and Chisago counties indicate a similar relationship as far west as Rush City and Rock Creek (Thiel, 1947, p. 19). In southeastern Kanabec County and northern Isanti County, drilling indicates the presence of both the Eau Claire and the Mt. Simon members over the Hinckley sandstone. This same relationship exists in the area of the Twin Cities and from there southward at the base of the Paleozoic rocks of the southeastern part of the State (Thiel, 1944, p. 60–61).

CORRELATION OF THESE BEDS WITH FORMATIONS IN WISCONSIN AND MICHIGAN: The Upper Keweenawan of northern Wisconsin and Michigan consists of a thick series of conglomerate, arkoses, sandstones, and shales. Thwaites (1912) has studied this series in northern Wisconsin and gives the following section:

Bayfield group

Chequamegon sandstone

	Feet
Red and white sandstone composed predominantly of quartz grains, with thin lenticular beds of sandy shale	1000

Devil's Island sandstone

Pink and white pure quartz sandstone with abundant ripple marks	300

Orienta sandstone

Like Chequamegon in its upper part, but containing more feldspar grains especially near the base	3000

From more recent field observations and from lithological and heavy-mineral evidence, Tyler *et al.* (1940, p. 1479) revised the sequence of Upper Keweenawan rocks in Wisconsin. This section has a total thickness of 17,850 feet, whereas Thwaites recorded a thickness of 24,000 feet.

Revised Section of Upper Keweenawan
(By Marsden and Tyler)

Bayfield group

Chequamegon sandstone	1000
Devil's Island sandstone	300
Orienta sandstone	3000

Oronto group

Freda sandstone	12 000
Nonesuch shale	350
Onter conglomerate	1200
	17 850

The correlation of the Hinckley sandstone and of the Fond du Lac beds with the various members of the Bayfield and Oronto groups of the Lake Superior sandstone of northern Wisconsin has not been established definitely. Atwater and Clement (1935) correlated the Hinckley sandstone with the Orienta formation of the Bayfield group, and the Fond du Lac beds with the Amnicon formation of the Oronto group, but the evidence for the correlation was not given. However, comparison of the accessory minerals of the Hinckley sandstone with those of various members of the Lake Superior sandstone as reported by Tyler *et al.* (1940) shows that the minerals of the Devil's Island sandstone of the Bayfield group are much like those of the Hinckley. Both contain very little garnet and no apatite. Zircon and tourmaline are the dominant accessory minerals in each. Likewise the accessory minerals of the Fond du Lac beds and those of the lower part of the

Orienta formation show similarities in mineral species. No sandstone equivalent to the Chequamegon sandstone, the uppermost member of the Lake Superior sandstone series, has been reported in Minnesota. If it was deposited, it has been removed by erosion.

Post Keweenawan Unconformity

The importance of the unconformity between the Upper Cambrian in Wisconsin and the whole Keweenawan group has been discussed by Atwater and Clement (1935).

Relative Importance of Unconformities

Five unconformities are listed in Table 3, but so little is known of the relations of the Sioux formation that a comparison between it and others is hardly feasible. The criteria of importance were listed by Blackwelder (1909, p. 289) and include (1) structure, especially discordance, (2) a missing series, or hiatus, and (3) a time gap. We have listed the pre-Knife Lake and pre-Animikie unconformities as great because the overlying beds are nearly normal to structures below, and enormous volumes of material are missing.

The post-Animikie pre-Keweenawan unconformity at several places shows very slight angular discordance. Blackwelder quotes Van Hise as referring to a "likeness of strike and dip," showing that this unconformity is "not great." A possible exception, which deserves further study, is the report (Van Hise and Leith, 1911, p. 215) of Keweenawan volcanic rocks lying on folded Cuyuna range rocks (Animikie).

The post-Keweenawan pre-Upper Cambrian unconformity is described by Atwater and Clement (1935, p. 2063–2066) as "large." Their feeling, based on careful surveys in Wisconsin, was that the break between these two was greater than would be expected if the Keweenawan were a continental deposit formed when marine sediments of Lower and Middle Cambrian were deposited elsewhere. Their conclusions are accepted, pending more radioactivity methods, as indicating that the Keweenawan is Precambrian. That does not mean, however, that the unconformity is "large" in comparison with the one before the Knife Lake group, or the one before the Animikie group.

Correlation

Introduction

Early discussions of the bases of correlation in Precambrian rocks were issued by Irving (1888). They were summarized and revised by Van Hise (1892); and again by Van Hise and Leith (1911) applied to the Lake Superior region. Later Leith (1934, p. 151–180) made some changes in the approved principles, chiefly that unconformities and the degree of metamorphism are not good bases for long-distance correlation. He then spoke hopefully of some new possible bases of correlation—radioactivity methods, identification of batholithic periods, the use of zircon types and possibly other heavy accessory minerals, petrographic features of iron-formations, and detailed areal mapping and metamorphic studies to show connections otherwise missed.

Leith's (1934, p. 169) recommendation that Precambrian rocks in a new area be described by "types", as Algonkian or Archean, has merit; but the purpose of this paper is to establish a succession, and that would be unnecessarily confused if parts of the Knife Lake group and parts of the Animikie are classed as of Archean type, while other parts are of Algonkian type.

The following paragraphs refer to earlier papers where discussions are satisfactory, add a few comments as to principles, and then suggest methods and apply them to Minnesota problems.

Principles and Methods

(1) Formations are correlated with other formations (or parts of formations). A formation is defined as a "mappable" unit, a "lithogenetic" unit; but by further study a unit may be subdivided into members that are mappable. Early formation names become names for a group of formations; it seems to be the path of progress to increase the number of units, but this should be done cautiously and conservatively. Formations commonly finger out, so that one formation at locality A may be equivalent to two or more at locality B.

(2) Correlation implies *time* equivalence, but two isolated formations may not always be *well* correlated (Schenck and Miller, 1941, p.

1423). In the absence of fossil evidence of age, radioactivity methods are about the only real method of estimating time. With progressively new and improved techniques, these are especially valuable in Precambrian igneous rocks.

Geologic time is continuous, and there are no intervals. Rastall (1944) says "Hard and fast limits for systems or even for eras are quite untenable, philosophically. Geologic time cannot be cut up into arbitary lengths." While mountains are built and eroded, sediments are being deposited elsewhere. Where a formation transgresses from one place to another and deposition occurs at a later time on one side than another, the principle of temporal transgression must be recognized (Wheeler and Beesley, 1947; 1948). Gaps in the sedimentary record are common, but *local*.

The time divisions should apply to some local type area, with time gaps at unconformities. If the local area has been carefully studied and the formations are widespread, it is more useful and valuable than a local area in which only a reconnaissance has been conducted and the systems are thin and in small areas. The area north of Lake Huron is recognized as an important type area.

Minnesota has a rather different sequence, but it has widespread groups, and their relations have been clearly determined. The Minnesota sequence is not yet well correlated with the Huronian.

The method of estimating time equivalence by "*walking the outcrops*" is widely used and seems serviceable if properly qualified (Cox *et al.*, 1921). Continuous outcrops of a formation, however, rarely extend for many miles. The question then arises as to how far interpolation can be carried across an overlying "surface" deposit. As the distances between outcrops increase and as the structures are complicated by faults, folds, and unconformities, such a method cannot safely trace a bed. Suggestive evidence may be derived from a characteristic *topographic expression* of a formation—it may be a "cliff-former", or "ridge-former"; or it may form gentle slopes or valleys. Certain beds may locally influence the *soil* and *vegetation* (or lack of it). Fragments of a formation in a hillside soil suggest that the formation lies up the slope. All the methods noted in this paragraph are of relatively little value in the glaciated area around Lake Superior.

Trenching, test pitting, and *drilling* have been applied locally to the problem of tracing a bed. They may be said to *make* an exposure where none was left by nature. Many beds have some peculiarity that can be detected by *geophysical or laboratory methods*, and, if these extend over a wide area, the bed can be traced very well. Even concealed structures such as faults and folds can at places be so clearly mapped that the offset beds can be well correlated. Samples must be taken if petrographic correlation is to be depended on, to test for accessory zircons, tourmalines, and other minerals, in the laboratory (Tyler *et al.*, 1940). Mass characters of insoluble residues and ratios of minerals may be diagnostic of certain beds. Fossils, commonly listed as occurring in Phanerozoic rocks, are so rare in the Precambrian that they are of little value. Gradual changes in a formation from place to place suggest the changes to be expected in a more distant exposure of the formation.

Key beds are of special value. The key bed may be thin or inconspicuous, but its value lies (1) in its persistence, and (2) in its peculiarity in a group of more ordinary rocks.

One Precambrian key bed deserves special mention as a basis for correlation. Leith (1934, p. 176–177) says that the iron-formations

> "are easily identified, are often well explored over extensive areas, and are known to be confined to few horizons. . . . New information makes it possible to correlate with fair degree of certainty the Negaunee iron-formation . . . with the Vulcan formation . . . with the Ironwood formation . . . with the Biwabik formation . . . and with the Animikie iron-formation. . . . This great . . . iron-formation now serves as a horizon marker to facilitate the correlation of sediments above and below."

The unique petrographic feature of this formation is the occurrence of a granule texture commonly attributed to a greenalite precipitate (Van Hise and Leith, 1911, p. 165–167, 521–552). The texture persists and is recognizable through a surprising amount of metamorphism and weathering, even when the minerals have wholly changed and, in some samples, apparently changed several times. Now such a tex-

ture resulting from a supposed precipitation of one or more minerals has no theoretical value—if it could develop in one geologic period, it might equally well result from similar processes in some other periods. Most such petrographic correlations are considered very questionable. In this particular example, however, there is a further line of evidence. The similarity of the granule textures in the iron-formations in several districts around Lake Superior was noticed very early and suggested a correlation. After close study and observation over a period of 100 years, it can be said that such textures have not been found in two formations of distinctly different periods in any district. It is on the basis of such experience, rather than on theoretical study of possibilities, that a texture in iron-formations becomes of great value in correlation. (Furthermore, the conclusions are supported and greatly strengthened by an internal sequence of key beds—discussed in Principle 3.)

The age of the Thomson formation is estimated in this paper on the basis of petrography, metamorphic structure, and the age of a granite that intrudes it. It is believed to be equivalent to the Knife Lake group. A similar problem comes up in the Cuyuna Range, but there the slates are tentatively correlated with the Animikie in spite of the fact that they are folded, faulted, and metamorphosed as much as the Knife Lake group. This apparent inconsistency is explained by the strength of the evidence of a key bed—the granule-textured iron-formation on the Cuyuna.

The conclusion from this long discussion is clear. A great deal more detailed mapping is needed to connect outcrops in rather widely separated areas. Leith (1934, p. 177) says

> "a better grasp of the time element in correlation will come naturally with clearer understanding of the geology of particular regions, and with the discovery here and there of local connections or near-connections between regions".

This is the path of progress in correlation.

Petrographic characters are properly used (Van Hise, 1892, p. 517) to correlate formations . . . in a single basin (or to distinguish dissimilar formations . . . when the older is most altered) only when they are not opposed to structural and stratigraphic evidence. They are suggestive but not conclusive.

Red beds occur at only a few periods in geologic time, and occurrence in adjoining districts suggests correlation.

Glasses in the Precambrian are more abundant in Keweenawan than in older formations, where most are devitrified.

Perthite is said (Faessler and Tremblay, 1946) to be more abundant and to have an ex-solution structure in late Precambrian, but to be less abundant and to have a replacement structure in earlier rocks.

(2A) The *origins* of rocks may at times be indicated or suggested by the petrographic detail. Rocks of the same origin are correlated with more confidence than those of different origin. Signs of different climatic environment during deposition (aqueous vs. eolian) discourage correlation. Origin in a small basin makes a sediment a questionable material for long-distance correlation. Tillites in two districts suggest correlation.

Some geologists believe that early Precambrian sediments were disintegrated without much decomposition, because organic acids were not abundant in soils. Archean sediments are characterized by graywacke rather than quartzite and limestone. Both quartzite and limestone first appear in the Huronian (Van Hise, 1909, p. 24) north of Lake Huron; but the supposedly older Grenville is not yet settled.

Sediments of geosynclinal origin are rarely correlated with "platform" sediments of the same district. It should, of course, be emphasized that sediments do form simultaneously in different environments.

(2B) *Chemical characters* may be listed as a special kind of petrographic property. For example, the fine clay matrix of a conglomerate may show the chemical effects of weathering, whereas a matrix of a tillite is unweathered rock flour. Moore (1918, p. 418) suggests cobalt is common in veins of Keweenawan age; but some is of other age too. Collins once thought a high potash content suggests Killarney granite; but this may have been based on incomplete studies. Wilson (1939b, p. 259) suggests Archean granite has mostly sodic feldspar; later granites are more potassic. Rankama

(1946) in a compilation finds younger rocks have more Li, Be, Rb, Cs, and Ba than older.

(3) *A sequence of beds* of several kinds is likely to be fairly uniform across one basin, or in a moderate-sized sedimentary province, but is likely to show differences at least in detail from the sequence in an adjoining basin. Correlations by sequences are greatly confused by local unconformities. Some sequences are common, and repeated in different basins, and in the same basin at other epochs. Other sequences, especially if they reverse the normal order, and especially if they include a *key bed* of exceptional peculiarity, are not likely to be repeated outside the basin and at other epochs.

The normal order in a marine sedimentary basin in one epoch is that of a transgressing sea. A rise of sea (or slow sinking of the crust), causing transgression, leaves a series of sediments buried and preserved under later sediments, commonly at progressively greater depths in the sea. In contrast to the transgression, a regression of the sea is commonly abrupt and associated with a more catastrophic rise of mountains. Any sediments deposited during that rise are liable to be promptly exposed to erosion which destroys the records. Many sedimentary series run (1) sandstone, (2) shale, and (3) limestone; relatively few show the reverse order.

"Only after (long) consideration of lateral changes in the facies distribution . . . produced by transgression and regression . . . can a rational reclassification be attempted" as an improvement on the common reference to sequences at type localities (McKee, 1945, p. 23).

There are in the Marquette district structural evidences of three series, each of which has (1) a basal quartzite, (2) a formation largely made up of precipitate from water, and (3) an upper slate. Evidently such a sequence could not be used for correlation, for the three sequences lie one above another.

On the other hand certain sequences in the members of the Biwabik iron-formation on the Mesabi range are so widespread and odd that a sequence having the same peculiarities in a district outside the Mesabi range can be correlated very certainly. Note the records (Wolff, 1917; Grout and Broderick, 1919; Gruner, 1924) including:

Upper slaty	Cherty limestone Thin-bedded taconite
Upper cherty	Septaria beds Hematite-magnetite bed Algal horizon Conglomerate cherty tacon (granule cherts)
Lower slaty	Thin-bedded taconite Intermediate slate
Lower cherty	White cherty and greenalite tac nite (granule cherts) Cherty magnetite taconite (gra ule cherts) Thin slaty layer Coarse algal pink chert

Such a sequence in the Gunflint formation i Canada is perfectly satisfactory as a basis fo correlation across the long gap where the Du luth gabbro cuts through the group so tha the beds cannot be traced.

(4) To establish a formation or a group on should determine not only its character but it boundaries above and below, and its relation to the overlying and underlying rocks. The may not be the same everywhere, but a grou unconformably above another in one district may have a conformable intermediate group i an adjacent district.

If a group shows only fault contacts at t type locality, it cannot be properly defined a all, and is difficult to correlate. Collins (1937 p. 1453) objected to the Timiskaming on thi basis. It is furthermore a poor basis for defini tion to locate groups only by reference to two other groups that are of very different age Cooke (1926) insists that it is not a good defini tion of Huronian to say that it lies "between the Archean and the Keweenawan". There is room in that interval for about 10 systems. (*See* Principle 11, below.)

(5) *Unconformities* are properly used (Van Hise, 1892, p. 518) to mark the boundaries of a group and correlate outcrops in a single basin; they may suggest correlation in separate basins, but caution is very necessary here.

The importance or magnitude of an unconformity is different at different places. Van Hise (1909, p. 37) says "the relative importance of unconformities can be determined where the

succession is full"; but the unconformity does not exist where the sediments were forming continuously in the full succession.

A recent critique of the time-stratigraphy concept (Wheeler and Beesley, 1948) says "rock units and unconformities may vary in age from place to place" so that "the employment of unconformities as time-stratigraphic boundaries should be abandoned". It may be maintained, however, that in the absence of better criteria, as is common in the Precambrian, the tracing of beds and unconformities is at least a basis for the suggestion of probable correlation.

Unconformities should be given extra consideration in the Precambrian (Alcock, 1934), because fossils are so rare, and because fossils, serving so well in the Phanerozoic, show time breaks at many unconformities. If districts, not too far apart, have similar unconformities and the same number of unconformities, there is a strong tendency to correlate the beds between the unconformities, and to correlate the formations below the lowest unconformity. Morley Wilson (1939b, p. 285–286) writes of tracing a "paleoplain" nearly 1000 miles. For this, one needs a series of *criteria* of unconformities.

Collins (1936, p. 1687) says if a conglomerate has *no granite* (granite suggesting deep erosion) but only pebbles of the underlying formation, it is probably local and not a good basis for correlation with another district.

Many attempts have been made to correlate beds by their relation to the supposedly equivalent unconformities in distant regions. Leith (1934, p. 167–174) says

"Better understanding of the nature and the significance of unconformities is rapidly undermining this ... prop to correlation. ... The surprising thing is that unconformities are as extensive and continuous as they appear to be. ... There is little evidence of simultaneous worldwide Precambrian folding. ... In limited areas, unconformities are useful, and often the most essential bases of correlation". "There is no known way to match them with any certainty in distant areas. They change in appearance, apparent magnitude, and number, in short distances."

The geologist attempting to use unconformities should study the valuation of unconformities (Blackwelder, 1909) in terms of discordance, hiatus, and duration of erosion. Men disagree as to the magnitude of unconformities. The time value of an unconformity involves regional study and a great deal of work. Even the great revolutions between eras, supposed to be worldwide, are not everywhere equally prominent.

(5A) It should be an acknowledged principle that no single group should be divided into different formations by a major unconformity; but minor disconformities are common between the formations in a group.

(6) Radioactivity methods of age determination can now be made accurate in favorable cases to 1 per cent. Even different methods now check within 5 per cent on favorable materials. In the Precambrian there is always a question whether the determined age is that of original formation or that of the latest strong recrystallization. Recent methods using pegmatitic lepidolite (Ahrens, 1947) and magnetite segregations (Hurley, 1943) are especially promising for the ages of large intrusives.

(7) Batholiths may be correlated with other batholiths. If batholithic invasions in the Precambrin are as widespread as in post-Cambrian time, and could be restricted (even in one continent) to three (or five) or any certain number of revolutions, the sequence of those invasions would give a good basis for correlating associated sediments.

The assumption of time equivalences of two batholiths in remote districts grew out of the theory of worldwide revolutions. (*See* Principle 5.) Granites without clearly related associated diastrophism (unconformity) should be used only with great caution. Holmes (1931) says that the radioactivity methods of age determination indicate at least seven times of great orogeny and granite intrusions in the Precambrian. Smaller intrusives are similarly scattered in several periods. The accessory zircon studies (Tyler *et al.*, 1940) show that the granites near Lake Superior are probably not all related to two simple invasions and revolutions (Table 2).

Not more than two periods of batholithic granite invasion should be expected to be evidenced by the rocks of a district. After a first granite, a second granite is likely to rise in the same area, and partly assimilate or obliterate

the first. A third is almost certain to make one of the earlier ones so obscure that its distinctness is left in doubt (Wilson, 1939b, p. 298).

Clearly the granite-batholith method of correlation must be used in connection with petrographic detail and ages determined by radioactivity methods.

(8) A series of other features give general suggestions as to relative ages of rocks in a district: *The older rocks have more deformation, tighter (isoclinal) folds, more recrystallization-metamorphism, and a larger proportion of igneous intrusives than the younger rocks in a sequence.* The degree or intensity of metamorphism may be measured in differing terms, one rock affected by dry heat, another by hot water, and a third by deformation, or some odd combination or succession of processes.

The method should be used only with great caution, and in small areas. Specifically, the post-Algoman Animikie iron-formation on the Mesabi range has gentle dips, but supposedly equivalent beds on the Cuyuna range and on the south side of Lake Superior are tightly folded and more metamorphosed. The Keweenawan flows are much more steeply tilted on the south side of Lake Superior than on the north. The Knife Lake slates with a low degree of metamorphism can be traced into schists and hornfelses of wide extent where influenced by intrusives.

(9) It may well be a general principle that a small outcrop of thin surface rocks is not as safe a basis for long-range correlation as thick extensive groups of sediments which indicate deposition in a wide basin. For illustration, the Knife Lake group has been considered equivalent to Timiskaming sediments farther east; but Timiskaming sediments are described as local deposits formed in small lakes where lavas temporarily dammed a valley until another flow buried them; whereas the Knife Lake sediments spread more than 200 miles along the international boundary and are enormously thick. Bruce (1927) found similar slates for hundreds of miles northwest and northeast, and they may occur outside the areas now known—formed in a real geosyncline.

(10) Where one or more groups in a succession of lavas, or sediments, or both, are found in the Precambrian *they should be given local names* until the evidence of their correlation with a standard or type locality is very clear.

(10A) In *naming formations*, priority has great strength, but long, well-established usage may at times prevail (Muller, 1941, p. 1431).

(11) Systems in the post-Cambrian probably represent 20 to 100 million years. Precambrian time is at least 1800 million years. This is long enough for 10 or 15 such systems (Schuchert and Dunbar, 1933, p. 72–79).

(12) A formation should be assigned to an age to indicate the time of its original deposition or solidification, not that of its metamorphism. Tanton writes (personal communication) of the Saganaga granite:

> "having diagnosed the big quartz grains . . . as post Ogishke, and since these are the very minerals that give the Saganaga granite its distinctive character, it seems to me *its age should be given as the date of the latest geological process identifiable as having operated in causing its formation*".

Probably the diagnosis is wrong, and *surely the conclusion is wrong*. If a Keewatin flow, a Knife Lake graywacke, and an Animikie slate could be traced to a hornfels at a contact of the Duluth gabbro, should all three be there assigned to a Keweenawan age? The answer is clearly, "No". The petrographic changes in a formation do not change the time at which it became a formation.

Results of these Methods

Since the Precambrian sequence in Minnesota has been established with only a few minor uncertainties, there is enough interest in the correlation with other districts to warrant a tabulation and brief discussion of problems. The correlation with near-by districts is much better than the correlation with places east of Lake Superior. Valuable attempts at correlation have been made by Leith, Lund, and Leith (1935) and by Pettijohn (1937). Other suggestions are included in the table, with only slight adjustments, which we hope do not distort the views of the authors materially (Seaman, 1943; Collins, 1937; Cooke, 1941).

Before long the radioactivity methods of determining age will probably be so improved that much better correlations will be possible.

Every effort should be made to encourage the researches on age by the use of magnetite and lepidolite in Precambrian igneous rocks. Until that is accomplished, we make the suggestions that are indicated by the data, and have faith that they serve well for areas near at hand. The Timiskaming and Grenville are too remote to justify even a guess, until radioactivity methods bring us new facts.

There is a noteworthy tendency, however, to extend the subdivisions downward as the upper (and simpler) groups become better known (Pettijohn, 1937). Cooke (1925) says,

> "the early Paleozoic was once considered as complex as the Archean"; "bit by bit the 'insoluble' problems have been solved"; "The historical outlook, therefore, leads us logically to believe that the conquest of the Archean is only a matter of time."

Possible Correlations

The *Keweenawan* correlations around Lake Superior are considered very satisfactory. They are based on tracing beds, similar petrographic materials in similar sequence, and relation to similar unconformities above and below; the only large granites cutting the series carry normal zircons suggesting Keweenawan age. We have not heard of any radioactivity test of the age of Keweenawan, there are no fossils, and there is only a little more deformation on the south shore than on the north shore of Lake Superior.

The very doubtful point in Keweenawan correlation is the suggestion that they may as a whole be continental deposits equivalent in age to Lower or Middle Cambrian, or both. Atwater and Clement (1935) working in Wisconsin conclude that the Keweenawan is Precambrian as here tabulated, but Seaman (1943) does not accept that for Michigan.

The Killarney granite at the type locality was reported by Collins (1925, p. 92) as late Precambrian, post-Huronian, and possibly cutting a series of Keweenawan dikes; in his latest paper (1937, p. 1446–1457) he reported that some granites near Killarney cut all the Bruce and Cobalt, Huronian sediments, but are of uncertain relation to the Whitewater beds and Keweenawan; for these he proposed the looser term, "Killarnean". In the absence of any study of the type of zircons and of any study of radioactive materials, correlation of the Keweenawan granites in Minnesota with the Killarney rocks is not attempted.

The *Sioux quartzite* (and very likely the Sibley formation in Thunder Bay) may be Keweenawan or between the Animikie and Keweenawan.

The *Animikie group*, named from an early description of rocks in the Animikie district, adjoining Minnesota on the northeast, is so close that it is almost possible to correlate with the type locality by "walking the beds." Only the Sibley formation of the Animikie district is missing in Minnesota.

Divisions of the Animikie were probably studied in detail on the Mesabi and Gunflint ranges, before detailed work was done elsewhere. We therefore retain the names, Pokegama quartzite, Biwabik iron-formation, and Virginia slate for the three major formations. Only minor breaks occur in this group in Minnesota.

The economic importance of these rocks is so great that many attempts at correlation have been made and opinions have changed as new evidence was discovered. By 1935 correlation of the Biwabik of Minnesota and the Ironwood of Wisconsin and the Negaunee of Michigan was considered well established (Leith, Lund, and Leith, 1935). The districts on the south shore of Lake Superior have a slate above this well-identified iron-formation with enough conglomerate between to suggest an unconformity. It was only natural then to search Minnesota for a corresponding unconformity. Up to the present time the evidences are poor and very inconclusive. On the west Mesabi there is, to be sure, some suggestion of an erosion and weathering of iron formation before the Virginia slate was deposited; but on close study it seems to be a minor break within the Biwabik formation—the rocks just above the break are thin-bedded iron-formation and not Virginia slate. Possibly, of course, deposition was continuous in Minnesota during the erosion noted south of Lake Superior. At least, nearly everyone agrees to the correlation of the iron-formations, but doubt may remain as to the correlation of the slates—which formations in Wisconsin, and Michigan are equivalent to the Virginia slate.

TABLE 4.—POSSIBLE CORRELATIONS

	Minnesota	North of Mesabi (Pettijohn)	Michigan (Two Seamans)	Marquette (L.L.&L.)	Michipicoten (L.L.&L.)	North of Lake Huron Type Locality (Cooke, 1941)
	Paleozoic		Paleozoic			
Preambrian (or Cryptozoic) Eon	Later: Keweenawan Group	Keweenawan		Keweenawan Acidic and basic intrusives	Basic dikes and flows	Killarney granite* Whitewater series, flows, sediments, and intrusives
			PROTEROZOIC: Sibley			
	Later: Animikie Group: Virginia, Biwabik, Pokegama	Animikie: Va, Bi, Po	Michigamme, Clarksburg, Bijiki I. Volcanics, Negaunee I. Volcanics, Abijik	ALGONKIAN TYPE — HURONIAN — Upper: Michigamme, Clarksburg, Greenwood I. F., Goodrich quartzite		Birch Lake granite*
				Middle: Negaunee I.F., Siamo slate, Abijik quartzite		Cobalt series — HURONIAN
				Lower: Wewe slate, Kona dolomite, Mesnard quartzite		Bruce series — HURONIAN
	Algoman granite	Algoman intrusive	EOZOIC: Algoman* granite		Algoman granite	granite*
	Medial: Knife Lake Group	Knife Lake sediments (and flows)	Kona dolomite, Mesnard quartzite, Kitchi		Dore conglomerate	
The	Pre-Knife Lake Granite	Granite intrusives	Laurentian granite	Laurentian* granite and gneiss	Granite & gneiss	Granite*

						conglomerate	
			Kitchi				

	Pre-Knife Lake Granite	Granite intrusives		Laurentian granite		Laurentian* granite and gneiss	Granite & gneiss	Granite*
	Earlier: Keewatin volcanics and Soudan formation	Kenoran volcanics Savant sediments	AZOIC		ARCHEAN TYPE	Kitchi	Helen I.F.	Sudbury series
		Granite intrusives						
		Keewatin volcanics Manigotagan sediments		Keewatin Mona schist		Mona schist	Greenstones and tuffs	Keewatin volcanics with sediments

* Doubtful age or position in the sequence.
– – – intrusive.
~~~ unconformity.
~~~

Seaman (1943) and Leith, Lund, and Leith (1935) take different positions on this problem. Gruner would question whether a thin lens of conglomerate at the top of a formation characterized by intraformational conglomerates is a sufficient basis for a claim that the overlying beds are unconformable.

The correlation of the Mesabi formations with the Cuyuna formations has been suggested many times, and, while the evidences in its favor are strong, several features leave it open to question. Disagreement is perhaps most serious as to the possible occurrence of iron-rich beds in the slate equivalent to the Virginia slate.

More remote correlation of the Animikie with the Huronian north of Lake Huron has been attempted many times, but is of very uncertain value. The logic of the argument runs: (1) The correlation is good from the Biwabik in the Mesabi district to the Negaunee in the Marquette district. (2) In the Marquette district this iron-bearing formation unconformably overlies a group (Mesnard quartzite, Kona dolomite, Wewe slate) which is fairly well correlated with the Bruce, Lower Huronian of the type locality (Collins, 1925, p. 114). (3) Above the Lower Huronian at Marquette are two unconformable groups, but above the Bruce of the type locality of the Huronian is only one, the Cobalt group, which was the original Upper Huronian. After two post-Lower Huronian groups were found at Marquette, a search was made for a second younger group north of Lake Huron. This resulted in a suggestion that the Whitewater rocks in the Sudbury basin might be the equivalent of the third group in Michigan. The Whitewater formation, however, is not well defined as to its relation to the Cobalt or to the Keweenawan. (4) In spite of this the suggestion was made that the Cobalt be called Middle Huronian, and the Whitewater Upper Huronian. Then there would be two groups above the Lower Huronian, both near Lake Huron and in the Marquette and adjacent districts. (5) On the strength of finding two possible groups between Lower Huronian and Keweenawan, correlation between the districts was assumed to be settled. Leith, Lund, and Leith (1935) correlate the Animikie with Middle and Upper Huronian.

This is not satisfactory, for the only basis for correlation of most Animikie iron ranges with middle and upper Huronian is that they lie between the Bruce (Lower Huronian) and the Keweenawan. As Lawson (1902, p. 57) pointed out, there is little petrographic resemblance between the Middle and Upper Huronian north of Lake Huron and the assumed Middle and Upper Huronian around Lake Superior. In the Lake Superior districts, iron-formations are continuous and similar for very great distances; but no such iron-bearing rocks occur in the Cobalt group north of Lake Huron. Other contrasts are equally striking. Collins (1925, p. 114) reports "no close resemblance". Cooke (1941) says "there has been nothing to fix the ages" of the Cobalt, Whitewater, and Animikie in relation to any Huronian except the Lower (Bruce). He considers the Whitewater as Keweenawan.

In an interval as long as that between the Lower Huronian and Keweenawan there is time for perhaps eight or more such groups, and there is no logic in the assumption that two groups in Michigan must be correlated with two different looking groups in Ontario. It is possible to make 28 different arrangements of groups 1 and 2 on the assumption of time enough for eight groups between Lower Huronian and Keweenawan. Is any one of the arrangements more probable than others?

A further reason to avoid the term Huronian for Animikie formations is based on the original naming of the Huronian. When Logan named the formations north of Lake Huron he thought (Alcock and Committee, 1934, p. 117) the Animikie and Keweenawan were Paleozoic and excluded them from Huronian.

Seaman (1943) does not use the term Huronian for the iron-bearing groups at Marquette but correlates the Animikie with the middle and upper Huronian.

The changes in original and later usage of the term Huronian, and the confusion as to what the terms now mean, led Miller and Knight (1913) to recommend dropping the term; A. Leith (1935) agrees.

Below the Animikie the structure is so much more complex than structure of most of the Animikie and Keweenawan that, based on Minnesota outcrops, the logical horizon for a

major unconformity is the base of the Animikie (Lawson, 1902). The lower great unconformity, at the base of the Knife Lake group, may be just as important, but it separates two groups segregations in related pegmatites. It appears to be of the same age as the Vermilion granite close to the north, which has been thus determined in preliminary and tentative work

TABLE 5.—POSSIBLE CORRELATIONS OF ANIMIKIE AND HURONIAN

Generally accepted correlation		Possible positions for placing two Animikie formations (1) and (2) in relation to Huronian
Keweenwan	K	
(Interval)		2 2 2 2 2 2 2 2 2 2 2 2
		2 2 2 2 2 2 2 2 2 2 2 1
Upper Huronian?	2	2 2 2 2 2 2 1 1
(Interval)		2 2 2 2 2 2 1 1 1
		2 2 2 2 1 1 1 1
Middle Huronian?	1	2 2 1 1 1 1
(Interval)		2 1 1 1 1 1 1
		1 1 1 1 1 1 1
Lower Huronian	LH	
(Interval)		
Pre-Huronian	Pre-H	

both of which are greatly deformed. Correlation is more difficult, and it may take longer to unravel the geology, but approximately two-thirds of geologic time is there represented.

No doubt the erosion before the Animikie deposition followed the exceptionally great intrusive and mountain-making activity of Algoman time. Seaman (1943) agrees that the pre-Animikie interval is long and important, but puts the Lower Huronian below that interval, thus dividing the type Huronian System by one of the greatest unconformities. There is serious question whether the great unconformity below the Animikie in Minnesota and Wisconsin is equivalent to (1) that below the Bruce series, (2) that above the Bruce series, or (3) some other.

The *Algoman* granite, with Giants Range of Minnesota as the type, is best correlated by radioactivity methods on a series of magnetite as about 750 million years old (Hurley, 1947). Mapping has shown that these granites are younger than the Knife Lake group and unconformably below the Animikie; and laboratory work shows no "normal" or Keweenawan type of zircons. These features indicate an age between the Knife Lake and Animikie groups, but leave an uncertainty as to correlation with granites on the south side of Lake Superior. (Table 2).

Ellsworth (1932) reported on the probable age of many granite pegmatite minerals which he studied for their lead-uranium ratios. By 1932 many samples had given tentative ages between 630 and 1200 million years; a distinctly older group gave ages near 1700 million years. Probably the Algoman was an important batholith period in the younger group.

The *Knife Lake* group is well correlated from the type locality near Knife Lake on the bound-

ary of Minnesota and Ontario, far to the west nearly to Lake of the Woods. It is altered from the typical slate to biotite schist near the intrusive Algoman granites. Correlation with the Thomson slate west of Duluth is less certain, but seems very probable.

Correlation outside of Minnesota is not very satisfactory, but the belt seems continuous from central Minnesota almost to Atikokan. Many other slate and schist belts to the north, northeast, and northwest are of the same type (Bruce, 1927), but their age (in relation to Animikie, Algoman, and older granites) is not so well established as that of the Knife Lake group.

Petrographically the Knife Lake group resembles Timiskaming more than it resembles Huronian. Wilson (1939b) says that his suggestion of this correlation was adopted by Miller and Knight (1913). We believe this correlation is an idea to work on, but for two reasons we prefer not to call the Minnesota rocks Timiskaming. (1) When the name Timiskaming was first proposed for sediments younger than Keewatin and older than Huronian, it was inferred that a granite had intruded the Keewatin and that both had been eroded profoundly so that the conglomerates rested on them. The geologists most experienced in the district (Wilson, 1939b, page 7), however, agree that the conglomerates have not been found resting on an eroded granite. The formation at the type locality is hypothetical (Collins, 1937). Several have suggested using the term Keewatin for flows, and Timiskaming for sediments *without age significance*. In Minnesota the Knife Lake group is much more definitely known, and its relations are established. It rests unconformably on Keewatin greenstone and on an old granite intrusive into that greenstone; it is intruded by Algoman granite; and both the Knife Lake and Algoman are overlain unconformably by the Animikie group. (2) Between the Algoman granite and the oldest granite, there is a time interval great enough to accommodate a great number of groups so that the Timiskaming may, as commonly stated (Cooke *et al.*, 1931; Collins, 1937), cover many local sediments at different horizons.

The *pre-Knife Lake granites* are ordinarily believed to be correlated with the Laurentian; but the type Laurentian locality is so far away, and the usage of the term Laurentian has been so varied that we use it only where it can be explained as equivalent to our "pre-Knife Lake." Aside from the Saganaga granite, there are some other granites along the north boundary of Minnesota which seem older than nearby Algoman granites. At Lake of the Woods a specimen of lepidolite (Ahrens, 1947) gave a tentative age of 2100 to 2350 million years. It is not yet certain that the Saganaga mass is closely correlated with the one that gave this estimate of extreme antiquity, but it is at least older than the Knife Lake sediments. North of the boundary Pettijohn (1937) suggests several sedimentary groups separated by more than one granite invasion and erosion older than the Knife Lake group.

The *Keewatin* having been named from outcrops on Lake of the Woods has its type locality in Minnesota, and the widespread use of the term is based on evidence (not all equally good) that other rocks can be correlated with those near Lake of the Woods.

It is possible that greenstones resembling those at the type locality may be of different ages. Small areas in the Knife Lake sediments are green andesitic flows; and Pettijohn (1937) reports greenstones of two ages north of the type locality. He would restrict the Keewatin to the older of the two series.

Cooke, James, and Mawdsley (1931, p. 52–53) suggest that there are larger, well-exposed, and less-metamorphosed greenstones in northern Ontario and Quebec, which have been so long agreed upon as Keewatin that it would be well to substitute that as the type locality. Nevertheless we retain the claim to the type locality, because there are good exposures in the Keewatin district from which the name comes.

Final Remarks

We have in Minnesota a rather detailed and exceptionally complete sequence of Precambrian rocks. If it is agreed upon, it may serve as a basis for rather extensive correlation with neighboring areas where the sequence is not so

extensive. Even the standard Huronian is not so satisfactory—its advantage is chiefly that the break at its base is a paleoplain that can be traced and well correlated for several hundred miles. Wilson's (1943) extensive sequence in western Quebec is so different and uncertain and so far away that correlation is doubtful.

If we include the area for 100 miles north of Minnesota we have fairly satisfactory evidence of granites of six different ages. The time is long from the early granites, about 2250 million years old, to Algoman granite, about 750 million years old (and Keweenawan even younger).

We would welcome a check on the relations of these formations by other geologists, in the hope that they can agree and establish the sequence as one of the most detailed anywhere recognized. From this area of detail it may then be possible to interpret other districts as corresponding to one part or another of this established sequence. Even then, it is entirely possible that groups (like the type Huronian) may be equivalent to gaps in the longer sequence here studied.

We definitely have in Minnesota a post-Keewatin granite of great age, which was eroded, and contributed fragments to a pre-Animikie conglomerate—that of the Knife Lake group. From this as a type locality it may be possible to correlate the similar rocks in an ever-widening area. While the existence of a post-Keewatin, pre-Timiskaming granite may be very doubtful as noted by Wilson (1939b, p. 289), and by Collins (1937, p. 1453), there is no doubt at all of the existence of a post-Keewatin, pre-Knife Lake granite at Saganaga Lake in Minnesota.

Several important problems still remain unsolved. (1) Correlation of the Minnesota groups with the Huronian groups is not certain. (2) It is uncertain whether the Keweenawan granites in Minnesota correspond at all closely with the Killarney granites at the type locality. (3) The relation of the Knife Lake group to the Timiskaming must await more study of the relation of some Timiskaming beds to older and younger rocks. (4) The Laurentian of the type locality needs a study by radioactivity methods to see if it corresponds to the old granites of the boundary area in Minnesota.

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MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, MAY 16, 1949.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 1079-1110, 13 FIGS. SEPTEMBER 1951

ENGINEERING GEOLOGY OF SAN FRANCISCO BAY, CALIFORNIA

BY PARKER D. TRASK AND JACK W. ROLSTON

ABSTRACT

The principal engineering problems that confront the geologist in San Francisco Bay are the strength of foundations, settlement of ground due to imposed load, stability of side slopes, support of piles, and source of borrow sand for fills. Work on the design of foundations for proposed new crossings of San Francisco Bay has shown that each geologic formation has more or less consistent characteristics with respect to these engineering problems. The geologist thus can facilitate the work of the engineer by advising him as to the engineering properties of the rocks and sediments in areas between bore holes or excavations, even though the bore holes may be several hundred feet apart.

Bedrock beneath San Francisco Bay consists of the Franciscan formation of Jurassic (?) age. Prior to the deposition of the present sediments in the bay, this formation was severely deformed, broken by faults, and eroded to a surface of considerable relief. The formation thus poses problems in landing caissons for bridges across the bay. In such places knowledge of the relief, firmness, and distribution of weak and strong rocks is essential.

The bay sediments consist of five formations of late Quaternary age, which correspond to deposits previously described by Lawson in his folio on the San Francisco Bay region. The oldest sediments belong to the Alameda formation, which ranges from 0 to more than 200 feet in thickness. The formation consists of a series of layers of firm sand, sandy clay, and clay, which everywhere forms a good foundation for engineering structures. The caissons in the east part of the bay on the present San Francisco-Oakland Bay Bridge rest on this formation. In more than 15 years the piers have not settled appreciably.

The overlying San Antonio formation, 15 to 120 feet thick, consists of clay and sandy clay, with a little sand. The sediments in most places are moderately firm and give good pile support. The formation settles appreciably under loads of more than 1.5 tons per square foot.

The succeeding formation, the Posey, 0 to 50 feet thick, consists of sand and sandy clay. It is firm and, where the underlying San Antonio formation is stiff, forms a good foundation. It also is a source of borrow sand. Subsequent to the deposition of the Posey formation, valleys 150 feet or more in depth and up to 1000 feet in width were formed. These valleys and adjacent flat land were then partially covered with the Merritt sand, 0 to 60 feet thick. This sand in part is wind-blown and in many places is a source of borrow sand.

The youngest formation is the Bay mud, which fills the old valleys and blankets the adjacent land. It is soft, affords a poor support for structures and piles, settles under imposed load, and fails readily when covered by significant quantities of fill. The formation causes critical problems in the deep, filled valleys. These valleys lie near the mouths of the present streams, which flow from the mountains. Presumably, the valleys mark the course of drainage when sea level was lower than it is now, perhaps during some glacial epoch.

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Introduction

The design of foundations for the proposed new crossings of San Francisco Bay involves a number of problems of engineering geology, such as the bearing capacity of sediments and rocks, settlement of sediments when covered by earth fill, stability of slopes of excavations and fills; and source of borrow sand for earth fills. These questions are closely related to soil mechanics, but, in applying the techniques of soil mechanics, the engineer is concerned with the extent to which he can extrapolate field and laboratory data beyond points of control. The geologist through his knowledge of the characteristics and behavior of earth materials can give quantitative information to the engineer as to whether the engineering properties of the rocks and sediments vary regularly or irregularly between or beyond the points of control. This paper describes the application of geology to some of these engineering problems in San Francisco Bay.

The work is based on an investigation of two proposed new crossings of the Bay: (1) a crossing essentially parallel to the present San Francisco-Oakland Bay Bridge, known as the Parallel Crossing, and (2) a crossing 2 to 4 miles to the south, consisting of a combination trestle, underwater tube, and mole causeway, known as the Southern Crossing (Fig. 1). The Southern Crossing extends from the south part of San Francisco to Alameda and is connected with Oakland by two parallel tubes beneath the Oakland Estuary. This proposed crossing of the Oakland Estuary is known as the Clay Street Crossing, in contrast with another proposed crossing, a quarter of a mile to the east and parallel to the present Posey Tube, which is called the Webster Street Crossing and is to be built in connection with the Parallel Crossing.

The engineering details of the two crossings have been described by Tudor (1948; 1949). Other publications relating to crossings of San Francisco Bay are indicated by asterisks in the bibliography.

The present study is based on field and

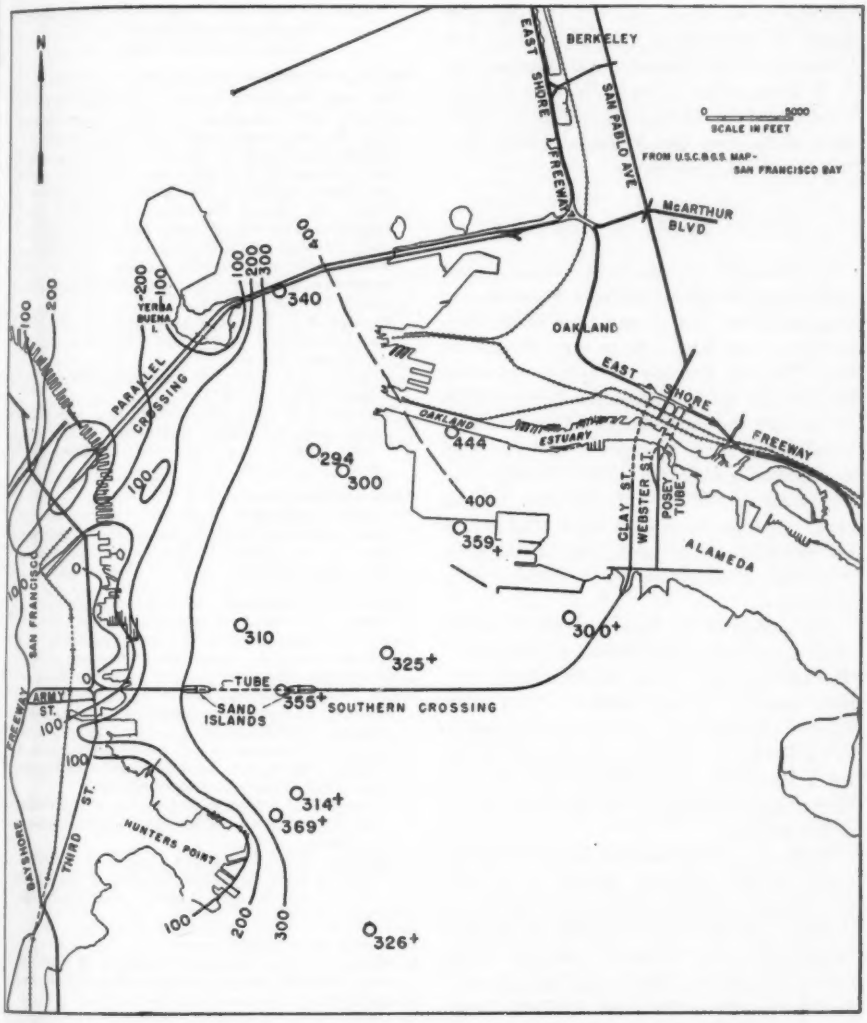

FIGURE 1.—GEOGRAPHIC FEATURES AND BEDROCK CONTOURS, SAN FRANCISCO BAY

Circles with adjacent numbers indicate location of borings and depth in feet to bedrock where bedrock is more than 300 feet below sea level.

laboratory investigations carried on between April 1948 and June 1950 by the Division of San Francisco Bay Toll Crossings of the Department of Public Works of the State of California. The following geologists and engineers were assigned to work on foundation problems: Karl L. Baumeister, Raymond F. Murphy, Don J. MacMullen, Edward J. Rhodes, Jack W. Rolston, Parker D. Trask, and John E. Wright. Trask was in charge of the work. Special acknowledgment is due these men for their help and co-operation and also to Ralph A. Tudor, Chief Engineer, and Norman C. Raab, Design Engineer, who were in administrative charge. The assistance of many other people is gratefully acknowledged: par-

ticularly O. R. Bosso, Louis J. Jennings, and Edwin F. Levy of the Division of Bay Toll Crossings; C. H. Harned, T. L. Stanton, and A. W. Root, of the California Division of Highways; and B. B. Gordon and Harold C. Shandrew of the East Bay Municipal Utility District, Oakland, California.

Geology of San Francisco Bay

San Francisco Bay has been a locus of earth movement throughout much of Mesozoic and Cenozoic time. The present form of the bay, however, dates back only to early Pleistocene or to Pliocene. The bay occupies a depression between two uplifted areas, the Berkeley Hills on the east and the Montara and other mountains on the west. Two active faults parallel the uplifted areas, each about 10 miles from the center of the bay: (1) the San Andreas fault, which passes through the central part of the mountains to the west; and (2) the Hayward fault, along the base of the Berkeley Hills to the east.

The geology of the region has been described by Lawson (1914) and Louderback (1939; 1951). The bedrock east of the Hayward fault is almost entirely Tertiary sandstone and shale. West of the fault most of the bedrock belongs to the Franciscan formation (Jurassic?), except for a small area of sandstone and shale, of Merced age (Pliocene), south of the city limits of San Francisco.

While the present bay was being formed the rocks on the adjoining uplifted areas were eroded, and a series of sedimentary formations deposited in the bay. According to Lawson (1914, p. 15) four formations crop out on the east side of the bay: Bay mud, Merritt sand, San Antonio formation, and Alameda formation. Lawson also mentions the Temescal formation, of terrestrial origin, but this formation is essentially equivalent in age to the Bay mud.

Each of these four formations is represented in the bay sediments. According to Lawson (1914), the San Antonio formation consists of a lower member containing no fragments of the Claremont chert (Miocene), and an upper member containing fragments of chert. This distribution of chert detritus leads Lawson to believe that the lower member was formed

Table 1.—Geologic Characteristics of Sediments and Bedrock, San Francisco Bay

Bay Mud—Formation A—Thickness 0-100 feet

Soft mud, becoming firmer with depth; consists of silty clay containing grains of wind-blown sand. In some areas upper part of formation includes thin layers of fine- to medium-grained sand alternating with silty clay. In places shells and plant fragments are found. Near Webster Street, the base of the formation is dark grey, suggestive of an old soil zone. The formation generally consists of three members. An erosional unconformity may overlie member A-3, as the upper two members are coarser in texture than zone A-3.

Member A-1 Soft mud composed of silty clay and occasional interbedded thin sand layers; thickness 0-65 feet.

Member A-2 Fine- to medium-grained silty sand; thickness 0-40 feet.

Member A-3 Silty clay, firmer and less sandy than zone A-1. Along Southern Crossing contains layer of peat near base; thickness 0-45 feet.

Erosional unconformity

Valleys attaining a depth of at least 150 feet below sea level, carved in underlying formations. Most of the valleys occupy valleys previously cut during post-Posey and pre-Merritt time.

Merritt sand—Formation B—Thickness 0-60 feet

Mostly well-sorted medium-grained sand, in part of wind-blown origin. The sand partially fills valleys carved in underlying Posey and older formations. The sand also lies as a blanket upon the Posey formation, and in places rises above sea level. In some areas upper part of formation contains admixtures of clay. In other places, part of the formation is well-sorted silt, suggestive of loess. The Merritt sand consists of two members, which locally are discontinuous.

Member B-1 Medium-grained sand containing varying amounts of silt and clay; thickness 0-20 feet.

Member B-2 Well-sorted, fine- to medium-grained sand, in places forming a permeable aquifer which may cause water problems; 0-60 feet thick.

Erosional unconformity

Valleys cut in underlying formations to depth of more than 150 feet below sea level and 1000 feet in width.

Posey formation—Formation C—Thickness 0-50 feet

Firm sandy clay and sand. In places contains rounded Franciscan pebbles up to an inch in diameter. The formation is firm and has been preconsolidated from 1.5 to 3 tons per square foot, presumably because of some process of desiccation. In some places, the Posey sediments are separable with difficulty from the Merritt sand, particularly east of Yerba Buena Island (Fig. 4). The formation consists of two members:

Member C-1 Firm sandy clay and clay; thickness 0-30 feet.

Member C-2 Fairly clean, fine- to medium-grained sand, in part seemingly of wind-blown origin; thickness 0-15 feet.

TABLE 1.—CONTINUED

Erosional unconformity
Erosion interval during which member D-1 and parts of member D-2 were removed.

San Antonio formation—Formation D—Thickness 15-120 feet
Moderately firm silty clay, in places containing thin lenses of fine gravel of Franciscan pebbles up to 1 inch in diameter. A persistent zone containing shells and sand is found near top of the formation in the east part of the bay. The lower part of the formation in many places is greenish gray, suggestive of glauconite. A black zone which seemingly is an ancient soil is found at the base of the formation near the Webster Street Tube. A more or less continuous layer of plant fragments lies a few feet above the base of the formation. The firmness of the sediments varies. In the vicinity of the Webster Street Tube, the deposits are relatively stiff, but east of Yerba Buena Island, where the formation is thickest, the sediments are soft. The formation can be separated into at least three members. The upper two members are not well developed in the west part of San Francisco Bay.

Member D-1 Silty Clay. Thickness 0-40 feet.

Member D-2 Persistent layer of shells associated with fine- to medium-grained sand and with silty clay. Not found on west side of San Francisco Bay. Thickness 0-40 feet.

Member D-3 Gray and greenish-gray clay. In places contains zones of sand or sandy gravel containing Franciscan pebbles. Persistent layer of plant fragments lies at base. Organic remains are found at other levels as well. Alternating layers of fine sand and clay are present. The clay in this formation is more fine-grained than clay in other formations. With additional bore holes this member probably could be subdivided further. Thickness 0-50 feet.

Erosional unconformity
Moderate erosion interval, during which perhaps as much as 50 feet of sediment was eroded from underlying formations.

Alameda formation—Formation E—Thickness 0 to more than 200 feet. Thickness varies with distance to bedrock.
Sand, sandy clay, clay, and fine gravel; commonly gray, but in places is greenish gray or brownish gray. Most of the gravel pebbles are less than 1 inch in diameter, well-rounded, and are derived from the Franciscan formation. Plant fragments are found at several horizons, particularly in upper part of formation. On west side of bay in hole 42 at a depth of 280 feet is a layer of clean white volcanic ash, relatively free from clastic material. This ash, is a dacitic vitric tuff (Max. D. Crittenden, Jr., personal communication). It contains 90 per cent glass and 10 per cent crystals, which include feldspar, quartz, and hornblende (?). The refractive index of the glass is about 1.52.
The Alameda formation contains at least five members of alternating sandy clay and sand. The sand commonly contains fine gravel. The lower part of the formation perhaps is equivalent to the Santa Clara or Merced formations.

TABLE 1.—CONTINUED

Erosional unconformity
Prior to deposition of the Alameda formation, the bedrock was deformed and eroded to a mature topography of considerable relief.

Bedrock—Franciscan formation (Jurassic?)—Thickness several thousand feet
In vicinity of the proposed crossings of San Francisco Bay, the bedrock consists of the Franciscan formation, composed of feldspathic sandstone, graywacke, siltstone, shale, interbedded chert and shale, and various types of basic and ultrabasic igneous rocks, now altered to greenstone and serpentine. The Franciscan has been broken by numerous small faults, which are no longer active. This formation is susceptible to landslides.

before the streams had dissected back to the outcrop of the Claremont chert in the Berkeley Hills. He therefore postulates that the main uplift of the Berkeley Hills is associated with the San Antonio formation, and hence the underlying Alameda formation antedates the main uplift.

The two members of the San Antonio formation cannot be directly identified in the bay sediments, but two lithologic units separated by an unconformity lie between the Merritt sand and the Alameda formation. These two units could represent the two members of the San Antonio formation described by Lawson, but since they are separated by an unconformity they probably are separate formations. The upper formation has been called the Posey formation, because it is well developed near the Posey Tube between Alameda and Oakland. The lower formation, ascribed to the San Antonio formation, consists of at least three members, which in part perhaps are the equivalent of the two members described by Lawson.

The principal characteristics of the five formations recognized in the San Francisco Bay are summarized in Tables 1 and 10. The individual formations can be traced fairly well from one bore hole to another. The correlation of the members within the formations is less certain, but most members are persistent, despite their terrestrial and brackish-water origin. With more bore holes and additional information the members probably could be separated further, but the classification in Table 1 has proved practicable in the present study.

Figures 2, 3, and 4 present profiles of the

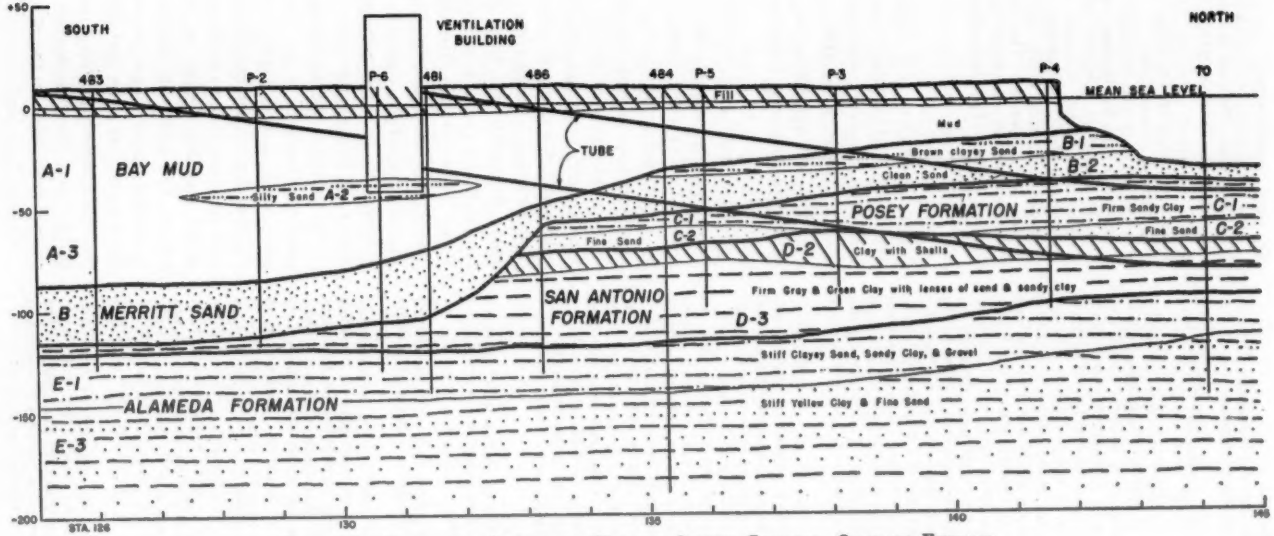

FIGURE 2.—PROFILE OF SEDIMENTS, WEBSTER STREET CROSSING, OAKLAND ESTUARY

Stations shown at base of drawing are 100 feet apart; vertical scale is in feet.

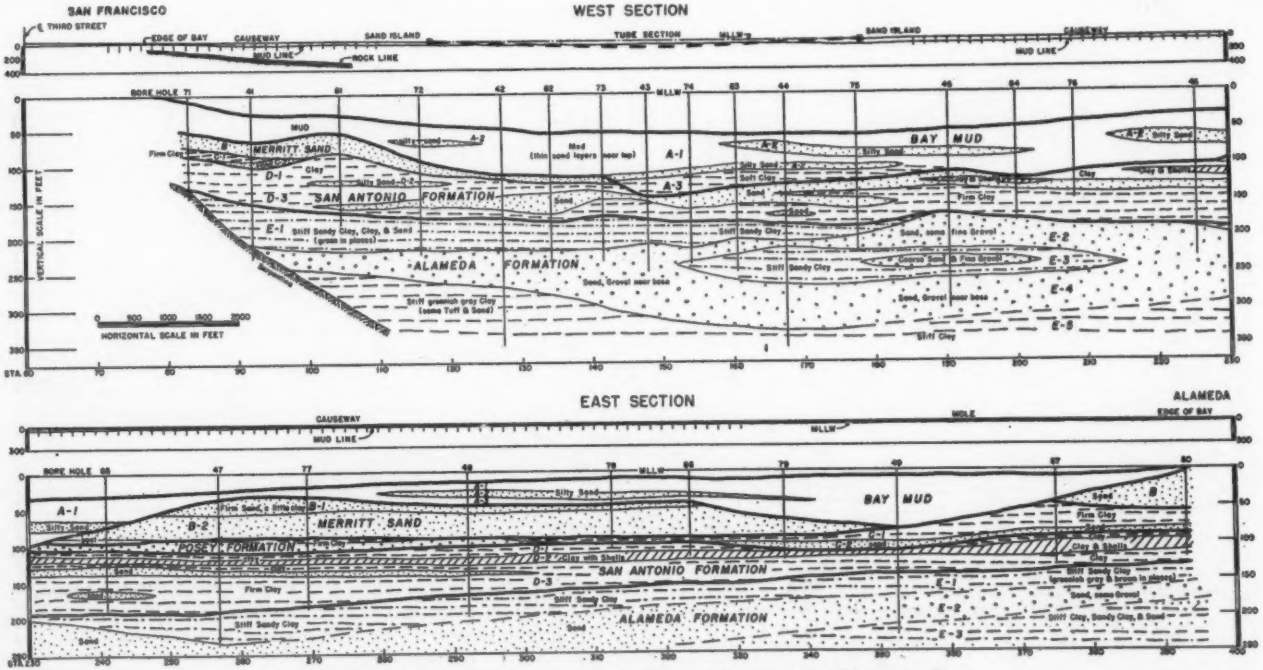

FIGURE 3.—PROFILE OF SEDIMENTS, SOUTHERN CROSSING, SAN FRANCISCO BAY

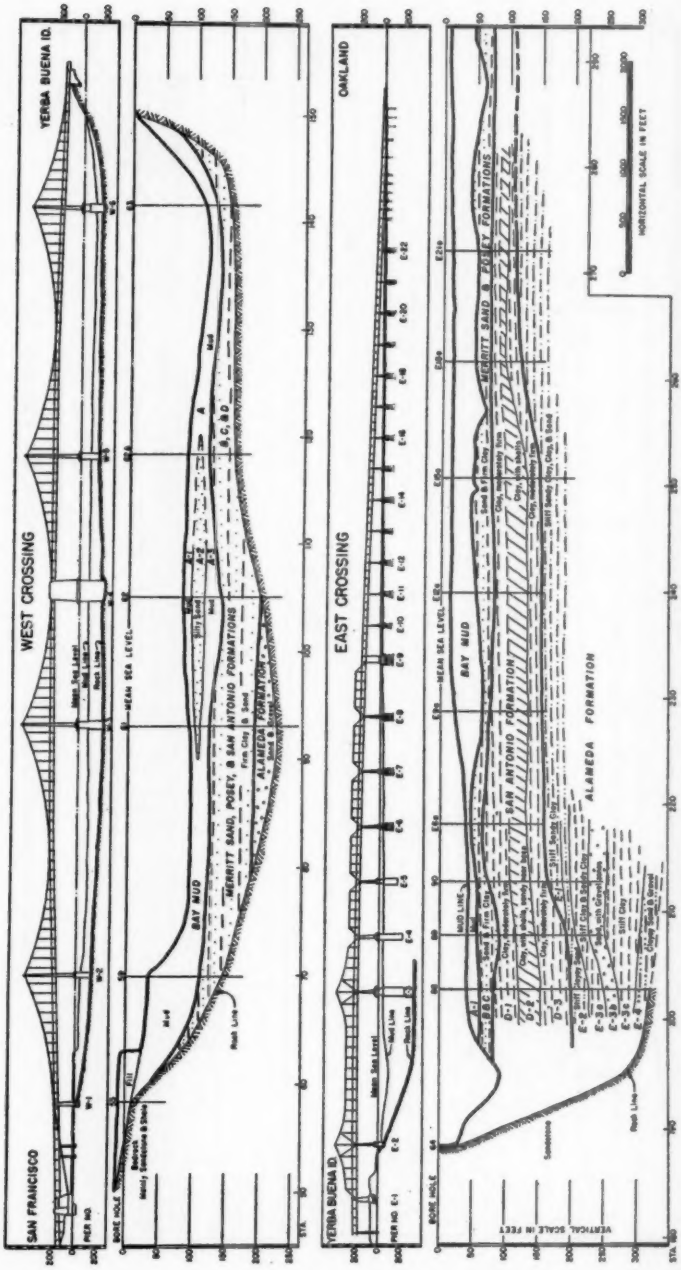

FIGURE 4.—PROFILE OF SEDIMENTS, PARALLEL CROSSING, SAN FRANCISCO BAY

TABLE 2.- LOG OF BORE HOLE 484, WEBSTER STREET, OAKLAND ESTUARY

(See Figure 2 for location)

Unit weight is reported as the average unit weight for the 1 to 9 sample tubes recovered from the sample interval. In general, unit weight for the respective tubes in a sample interval is essentially constant; but where the range between maximum and minimum unit weight is 6 or more pounds per cubic foot, the symbol x is used. Jf refers to Franciscan formation.

Elevation	Sample Length	Sample No.	Geol. Zone	Unit Wt.	Color	Remarks
+6.0-+3.0	1.7	1	Fill	122	Gray	Fill (quarry run - sandstone & clay)
+1.0--2.0	0	2	Fill	-	-	No Recovery
-2.0--6.0	2.7	3	Fill A-1	- 94	Gray Gray	Fill (round medium sand) Mud (silty clay - trace of organic matter)
-11.0-14.0	2.3	4	A-1	93	Gray	Mud (silty clay, few rd. med. sand grains — little organic matter; few shells
-21.0-24.0	2.3	5	A-1	99	Gray	Mud (silty clay, few rd. med. sand grains — few shells; Foraminifera
-31.0-34.0	2.3	6	A-1 B-1	103 118	Gray Gray	Mud (soft silty clay, few round med. sd. grains Medium fine round silty sand
-36.0-39.0	1.7	7	B-2	134	Gray	Medium fine round sand, little silt
-47.0-50.0	1.7	8	B-2 B-2	128 130	Gray Grn-gray	Fine silty sand Firm clay, few coarse round sand grains
-52.0-55.0	1.0	9	B-2 B-2	125	Gray Gray	Round medium silty sand Firm silty clay; few round med. sand grains
-57.0-60.0	2.7	10	C-1 C-1 C-1	- - 127	Grn-gray Grn-gray Grn-gray	Firm medium sandy clay Round medium fine silty sand Firm silty clay, few round sand grains
-62.0-65.0	2.0	11	C-2	130	Gray	Round med. fine silty sand, little clay; few shells
-67.0-70.0	2.0	12	C-2 C-2	120	Grn-gray Grn-gray	Round medium silty sand Firm medium fine sandy clay
-72.0-75.0	3.0	13	D-2	113	Gray	Silty clay & sand; round fine sand grains; many oyster shells
-77.0-80.0	2.7	14	D-2 D-3	123 112	Gray Gray	Silty clay & fine sand; shells Silty clay, little round medium sand
-82.0-85.0	2.3	15	D-3	104	Gray	Silty clay, little round medium sand — plant fragments
-87.0-90.0	0	16	D-3	-	-	Lost sample
-90.0-93.0	3.0	17	D-3 D-3	129x 125	Gray Grn-gray	Medium fine silty sand; clay matrix Firm silty clay, medium fine sand
-95.0-98.0	1.0	18	D-3	117	Grn-gray	Firm clay; few round medium sand grains
-100.0-103.0	2.7	19	D-3	112	Grn-gray	Silty clay; few round medium sand grains — little organic matter
-105.0-108.0	1.3	20	D-3	108	Grn-gray	Silty clay; few round medium sand grains — little organic matter; few, Jf pebbles to 1/4"
-110.0-113.0	2.0	21	D-3	110	Gray	Clay; many plant fragments; few shells at base
-115.0-118.0	0.3	22	D-3	-	Gray	Silty clay, med. fine sand; tr. organic matter
-120.0-123.0	2.7	23	E-1 E-1	130x	Gray Gray	Medium coarse sandy clay; few Jf pebbles to 1/4"; organic matter Silty clay, fine sand; organic matter
-123.0-128.0	1.7	24	E-1 E-1	129	Gray Grn-gray	Medium fine silty sand; few round grains Medium fine sandy clay
-130.0-133.0	3.0	25	E-2 E-2	136 -	Gray Gray	Medium fine silty sand & clay Coarse round sand & fine gravel, little clay; Jf pebbles to 1"
-133.0-134.0	0	26	E-2	-	-	No Recovery
-156.0-160.0	2.0	27	E-3	121 to 133	Yel-brn.	Silty clay; few rd. med. sand grains; few plants
-165.0-168.0	2.3	28	E-3	132	Yel-brn.	Medium fine sandy clay
-173.0-176.0	2.3	29	E-3	131	Yel-brn.	Fine sandy clay; few plants
-181.0-184.0	2.3	30	E-3	129	Yel-brn.	Fine silty clay, little fine sand; few plants

ZONE DEPTH
48
A-1
73
A-3
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B 117
D-1 125
D3-a 140
D3-b 160
D3-c 174
E-1
219
E-2
240
E-4
270
E-5
350
HOLE

sediments on the Webster Street, Southern, and Parallel crossings respectively. Typical logs for these three crossings are given in Tables 2, 3, and 4, and in Figure 5. The Webster Street Crossing is described first, because the stratigraphic relationships are more clearly indicated in this area.

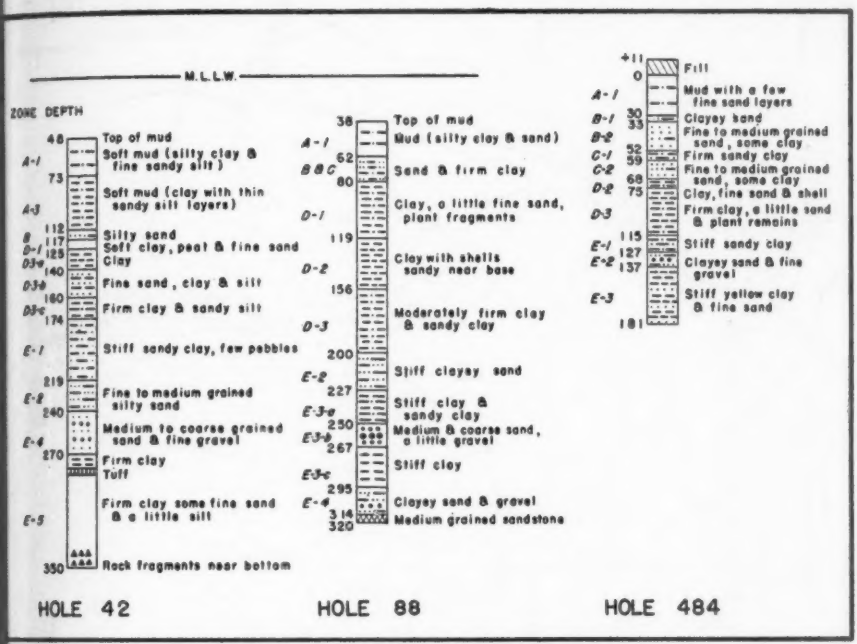

FIGURE 5.—LOGS OF TYPICAL BORE HOLES, SAN FRANCISCO BAY
Depth is in feet below mean lower low water.

Each formation seemingly is separated by an unconformity, the most pronounced of which lies between the Posey and Merritt formations. During the time between the deposition of these two formations, valleys were cut to a depth of at least 150 feet beneath sea level (Fig. 6). Each valley lies near the mouth of one of the present streams flowing westward from the Berkeley Hills. Along the present shore line of the east side of the bay, the bottoms of the valleys are 30 to 100 feet below sea level. These valleys drain into a master valley that extends northward across the submerged rock ridge west of Yerba Buena Island. This master valley is more than 150 feet below sea level where last observed west of Yerba Buena Island.

The valleys in part are filled with Merritt sand, which in turn was subjected to erosion prior to the deposition of the Bay mud. In places, the valleys formed during these two erosion periods are 1000 feet in width. They could well be associated with glacial lowering of sea level.

The unconformity between the Posey and San Antonio formations is slight, but in places the upper part of the San Antonio formation (member D-1) seemingly has been cut away. The unconformity between the San Antonio and Alameda formations may be of considerable magnitude, because along the west part of the Southern Crossing (Fig. 3) member E-1, which is more than 50 feet thick, apparently has been removed by erosion.

The contact between the San Antonio and Alameda formations slopes toward the center of the bay approximately 25 feet per mile. The

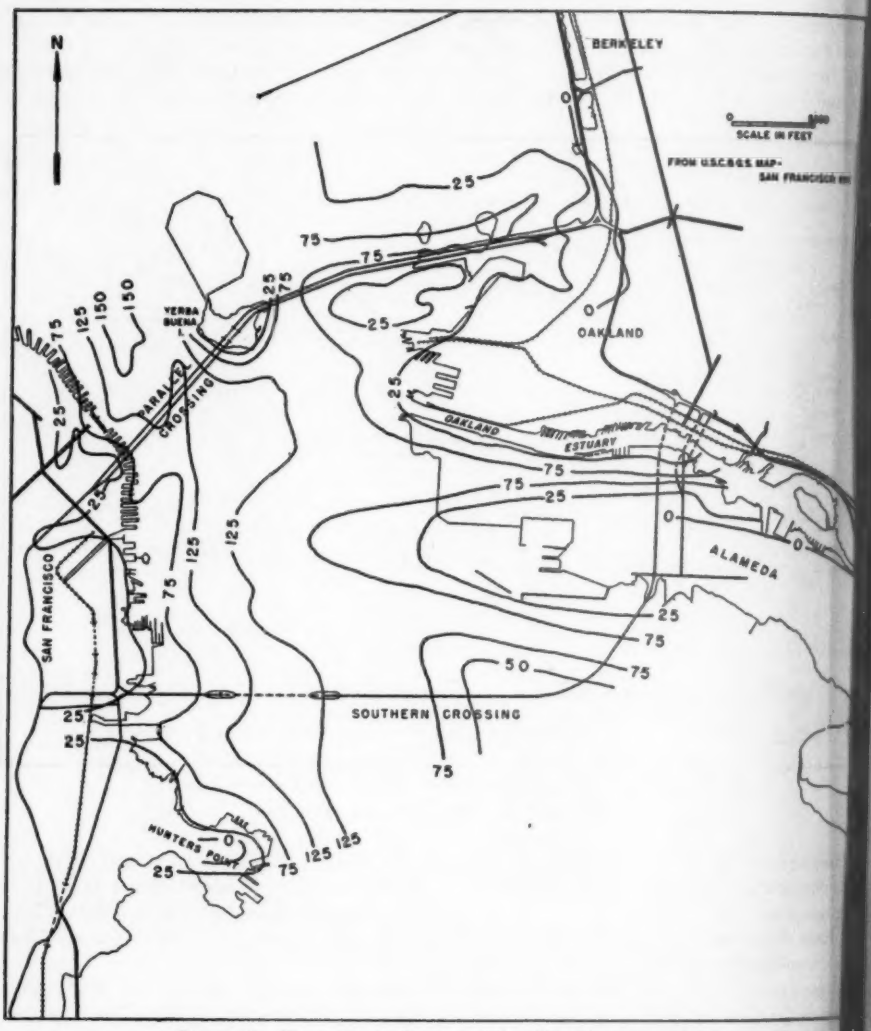

FIGURE 6.—CONTOURS ON BASE OF MUD, SAN FRANCISCO BAY
Contours are in feet below mean lower low water. Note irregular interval.

stratigraphic relationships of the formations suggest a more gentle slope of deposition; consequently, the central part of the bay may have been downwarped since deposition of the Alameda formation.

The Alameda formation as classified in this report includes all the soft sedimentary deposits between the San Antonio formation and Franciscan bedrock. Part of the Merced (Pliocene) and Santa Clara (Pliocene and Pleistocen formations may be included in these sedimen The master stream, extending northward acr the bedrock at the base of the Alameda for tion, flowed east of Yerba Buena Island (Fig. Bedrock in this area is now at a depth of ab 325 feet. The water at the deepest point in present channel in the Golden Gate is appro mately 360 feet deep. As the strong tidal sc

TABLE 3.- LOG OF BORE HOLE 42, SOUTHERN CROSSING, SAN FRANCISCO BAY

(See Figure 3 for location and Table 2 for explanation.)

Elevation	Sample Length	Sample No.	Geol. Zone	Unit Wt.	Color	Remarks
-48.3-53.3	2.7	1	A-1	94	Gray	Soft mud (silty clay)
-59.0-63.0	1.7	2	A-1	95[x]	Gray	Soft mud (silty clay), few round medium sand grains; shells
-66.3-70.3	3.0	3	A-1	99	Gray	Soft mud (fine sandy silt), few round med. sand grains; few shells
-74.7-77.7	3.0	4	A-2-b	96	Gray	Soft sticky mud (clay, with thin fine sandy silt layers); organic matter
-82.0-85.0	2.7	5	A-2-b	94	Gray	Soft sticky mud (clay, with thin fine sandy silt layers)
-89.0-92.0	3.0	6	A-2-b	80 to 99	Gray	Soft mud (clay, with thin fine sandy silt layers)
-96.7-99.7	3.0	7	A-2-b	97[x]	Gray	Soft mud (clay, with thin fine sandy silt layers); shell fragments
-104.3-107.3	3.0	8	A-2-b	98	Gray	Soft mud (clay, with alternating sandy silt & silty clay layers
-112.3-114.8	2.5	9	B	102[x]	Gray	Soft silty sand; round grains; few shells
-119.7-122.2	1.7	10	D-1 D-1	76 101[x]	Gray Gray	Soft clay, with peat Soft clay, with fine sand & plant fragments
-127.4-129.9	2.5	11	D-3-a	104[x]	Gray	Soft clay, with plant fragments
-135.6-138.1	2.5	12	D-3-a	98 to 110	Gray	Soft clay; trace of organic matter
-143.7-146.2	2.5	13	D-3-b	125	Gray	Fine sand, with soft clay layers; round grains
-151.5-154.0	2.5	14	D-3-b	123	Gray	Fine sand, some silt; round grains
-159.1-161.6	2.5	15	D-3-b D-3-c	114	Gray Gray	Thin layers of silty sand & silty clay Fine sandy silt
-167.7-170.2	2.3	16	D-3-c	119	Grn-gray	Silty clay; plant fragments
-174.8-177.3	2.5	17	E-1	122	Grn-gray	Clay, with few rounded Jf pebbles to 1/4"
-181.7-184.2	2.5	18	E-1	126	Grn-gray	Fine sandy clay
-188.7-191.2	1.7	19	E-1	124	Grn-gray	Very sandy clay
-195.7-198.2	2.0	20	E-1	122	Grn-gray	Fine sand with silt (disturbed sample)
202.8-205.3	2.0	21	E-1	124[x]	Grn-gray	Medium coarse sand & clay layers
209.4-211.9	2.3	22	E-1	126	Grn-gray	Silty clay; pebbles to 1/4"; Foraminifera?
216.8-219.3	2.5	23	E-1 E-2	126 124	Grn-gray Grn-gray	Fine silty sand, with few round serpentine pebbles to 1/2"; organic matter Fine silty sand
224.8-227.3	1.7	24	E-2	129[x]	Grn-gray	Fine silty sand
233.1-235.6	1.7	25	E-2	127	Grn-gray	Med. sand & some silt; round grains
241.6-244.1	2.0	26	E-4 E-4	130	Grn-gray Grn-gray	Med. sand & some silt Med. sand; clean, rd. grains
251.4-254.5	0	27	E-4	-	Grn-gray	Lost sample - Wash sample shows coarse sd. & gravel, Jf peb. to 1/4"; slightly rounded
254.5-257.0	1.7	28	E-4	138[x]	Grn-gray	Gravel, coarse sand & silt; pebbles to 1"
263.2-265.7	2.0	29	E-4	138[x]	Grn-gray	Coarse sand & gravel, to 1"; pebbles well rounded, clean
271.8-274.3	2.3	30	E-5	134	Gray-grn	Firm sandy clay
279.8-280.8	1.0	31	E-5	103	White	Tuff (volcanic ash)
282.8-285.3	2.5	32	E-5	127	Green	Firm clay
-287±	-	-	E-5	-	Grn-gray	Boiler sample - clay with shells
290.0-292.5	2.3	33	E-5	130	Brn-gray	Fine silty sand
297.0-299.5	2.5	34	E-5	126	Gray	Stiff silty clay, few round medium sand grains
300.0-302.5 (Approx.)	-	-	E-5	-	-	Bit sample (not saved) - fine sand, silt & clay
302.8-305.3	2.5	35	E-5 E-5	129	Grn-gray Gray	Fine silty sand Stiff clay
310.5-313.0	2.3	36	E-5	128	Gray	Clay, trace of sand
318.0-320.5	2.5	37	E-5	123	Gray	Silty clay; Foraminifera?
323.8-326.3	2.5	38	E-5	126	Grn-gray	Firm clay; Foraminifera?
329.9-332.4	2.0	39	E-5	126[x]	Brn-gray	Firm clay
335.9-338.4	2.3	40	E-5	124	Gray	Firm clay
343.2-346.2	2.3	41	E-5	122	Gray	Firm clay; rock fragments to 1/4"
347.8-350.3	2.3	42	E-5 E-5	114[x] 112	Grn-gray Light-gray	Firm clay; rock pebbles to 1/4", rounded Clay, friable, (tuff?); many rock chips

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in the Golden Gate may have eroded all the soft sediments, this depth perhaps is a measure of the maximum depth of the old valley in San Francisco Bay, disregarding of course subsequent diastrophism. As bedrock is encountered at progressively greater depths eastward from Yerba Buena Island, being more than 400 feet deep in the west part of the city of Alameda (Fig. 1) and deeper eastward, it would seem to follow that, prior to the time that a river last flowed east of Yerba Buena Island, sediments had been deposited on the bedrock to the east. Thus presumably an unconformity of considerable magnitude lies within the beds correlated with the Alameda formation. The sediments beneath such an unconformity, however, are so deeply buried that they have little effect upon engineering problems in the construction of bridges across San Francisco Bay.

Basic Data Procured

Bore Holes

The present investigation is based primarily on 60 bore holes drilled in 1948 and 1949 along the sites of the proposed crossings of San Francisco Bay (Figs. 2, 3, 4). The samples procured from these bore holes were tested in the field and in the laboratory according to present methods of soil mechanics.

Additional information is derived from bore holes drilled during previous investigations for engineering construction in San Francisco Bay, principally: Hoover-Young report (1930, p. 165–193), California Department of Public Works (1947, p. 165–175), and Whitworth (1932). Other bore-hole data were supplied by local organizations, individuals, and government agencies. Maps showing the location of all drill holes are on file in the office of the Division of San Francisco Bay Toll Crossings in Berkeley, California.

The samples obtained during the boring program of the Division of San Francisco Bay Toll Crossings in 1948 and 1949 were taken with a California sampler in open and in closed positions (Stanton, 1936, p. 13), and with a Shelby open and a Shelby piston sampler (Parsons, 1948, p. 63–68). The California sampler delivers samples in thin-walled brass cylinders 2 inches in diameter and 4 inches in length. The Shelby samplers give samples 12 to 18 inches in length and approximately 3 inches in diameter. In operating open samplers, the drill hole was cased and cleaned to the depth at which the sample was taken. When the California sampler was used in the open position, it was driven by means of jars weighing approximately 1750 pounds and having a stroke of 2 feet.

In the 1948 operations (Holes 41 to 90), 6336 feet of sediments was drilled, of which 1710 feet was recovered from 859 samples, representing a ratio of sampling to drilling of 27 per cent. Twenty-two samples were lost. The depth of holes ranges from 100 to 350 feet. Three holes (Nos. 41, 71, 88) encountered bedrock. The others terminated in sediments.

In the 1949 operations along the Webster Street Crossing (Fig. 2), six holes (Nos. 481–486) aggregating 904 feet in length were drilled, from which 325 feet, or 36 per cent, was recovered as samples.

The rock cores were taken in holes 50–64 by means of a double-tube, ball-bearing, swivel-type barrel, using an NX diamond bit. The cores were slightly more than 2 inches in diameter. Of 1123 feet of rock cored, 604 feet or 54 per cent was recovered. Two holes at the west anchorage and three near Pier W-1 (Fig. 4) drilled in nearly vertical thin-bedded sandstone, shale, and chert, and in greenstone, gave a recovery of 25 per cent; six holes beneath the water in similar types of rock yielded 30 per cent; and massive sandstone and siltstone on Yerba Buena Island gave 80 per cent recovery.

Analyses

General statement.—The sediments and rock cores were examined in the field and in the office. The usual soil-mechanics tests were made in the laboratory of the Materials and Testing Laboratory of the California Division of Highways in Sacramento under the direction of T. E. Stanton and A. W. Root.

The field examination of the samples included classification of the texture and the determination of the wet unit weight (lump density) of the samples as recovered from the sampling instrument. Each sample also was examined microscopically in the office.

Laboratory tests were made as follows: moisture content, 686 tests; mechanical analyses, 696; Atterberg limits, 463; specific gravity, 50; ignition loss, 19; consolidation, 265; unconfined compression, 565; direct shear, 86; and consolidated quick shear, 10 tests. In addition, triaxial shear tests were performed on seven samples from the Southern Crossing by the O. J. Porter Company of Sacramento, and nine Shelby samples from the Oakland Estuary (Holes 481 and 482) were tested in detail in the laboratory of Moran, Proctor, Freeman, and Mueser of New York, in order to determine the degree of consolidation of the soft Bay mud previously covered with earth fill. The results of these analyses are on file in the Division of San Francisco Bay Toll Crossings. The laboratory data are summarized here.

Unit weight.—The unit weight is the weight of a given volume of sediment, expressed in pounds per cubic foot. It corresponds to the lump or bulk density. In the present work the 4-inch brass sample tubes filled with sediment were taken from the sampler, the ends squared off, the tubes weighed, and the weight converted to pounds per cubic foot. More than 3000 determinations were made in the field. The unit weight proved valuable in correlating the sediments, as well as for indicating the general engineering properties.

The unit weight of saturated sediments is an index of the water content, as it represents the weight of the water plus the weight of the solid constituents of a given volume of sediment. As discussed in the section on Engineering Applications of Geology, the water content is an index of the strength of the sediments. Deposits with high water content have little strength compared with those of low water content.

The water content of sediments saturated with water, such as those in San Francisco Bay, can be computed from the unit weight, provided the density of the detrital particles is essentially constant. The results of specific-gravity determinations of 50 representative samples from the bay indicate that the density of the solid constituents deviates relatively little from 2.66.

The basic formula for using the unit weight to calculate the water content of saturated sediments is

$$w = \frac{d_1(d_2 - d_s)}{d_2(d_s - d_1)}$$

where w is the ratio of the weight of the water to the weight of the solid constituents (not the ratio of water to total weight of the sediment), d_1 is the density of the water, d_2 the density of the solid constituents, and d_s the bulk density of the sediment. As the unit weight is expressed in pounds per cubic foot and water of specific gravity 1.00 weighs approximately 62.5 pounds per cubic foot, the water content W of the sediment is determined by the formula

$$W = \frac{100(166 - \gamma)}{2.66\gamma - 166}$$

where W is the water content expressed as per cent (100 w in the previous equation), γ is the unit weight in pounds per cubic foot, and the specific gravity of the water and solid constituents is 1.00 and 2.66, respectively. The brackish water of the bay is slightly denser than fresh water but the difference is not sufficient to make an appreciable difference in the calculations.

The distribution of unit weight among the formations in San Francisco Bay is shown in Table 5 in which the unit weight is averaged for each formation for each of six areas, and also for all areas.

The sand in San Francisco Bay as a rule has a greater unit weight than the clay and silt. Also, the unit weight increases more or less progressively with increasing depth, owing mainly to the expulsion of water while the sediments are compacting. The most notable exceptions to this generalization are the Posa and San Antonio formations, which have been preconsolidated, presumably during the pre-Merritt erosion interval. The clayey sediments in the upper part of the Bay mud in the east part of the Southern Crossing have a higher unit weight than the corresponding sediments to the west, probably because the sediments are coarser.

Though the average unit weight of the individual formations is fairly constant throughout the areal extent of each formation, the un-

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Though the average unit weight of the individual formations is fairly constant throughout the areal extent of each formation, the unit

TABLE 4,- LOG OF BORE HOLE 88, PARALLEL CROSSING, EA

(See Figure 4 for location and Table 2 for e

Elevation	Sample Length	Sample No.	Geol. Zone	Unit Wt.	Color	
-50.3-53.3	2.3	1	A-1	95	Gray	Mud (si
-58.9-61.9	2.3	2	A-1	112 to 93	Gray	Mud (si
-66.9-69.9	2.0	3	BC	130	Black	Medium
-74.8-77.8	3.0	4	BC BC	124 117	Gray-grn. Gray-grn.	Sandy c Clay
-82.8-85.8	2.3	5	D-1 D-1	106 85	Gray Dark-gray	Silty cla Clay, lit
-90.6-93.6	2.3	6	D-1 D-1	119 106[x]	Gray-grn. Gray-grn.	Firm sa Clay, lit
-98.2-101.2	2.7	7	D-1	104	Gray	Silty cla
-106.1-109.1	2.7	8	D-1	102	Gray	} Silty c
-113.7-116.7	2.7	9	D-1	104	Gray	
-121.5-124.5	2.7	10	D-2	102	Gray	} Silty c
-129.2-132.2	2.7	11	D-2 D-2	103	Gray Gray	} Silty c
-136.8-139.8	2.7	12	D-2 D-2	105	Gray Gray	Silty cla
-144.5-147.5	2.7	13	D-2	107	Gray	Silty cla
-152.3-155.3	2.3	14	D-2	105	Gray	Silty cla
-159.8-162.8	2.3	15	D-3	119	Grn-gray	Sandy c
-167.5-170.5	2.7	16	D-3	107	Grn-gray	} Clay, l
-175.1-178.1	2.7	17	D-3	104	Gray	
-182.7-185.7	2.7	18	D-3	103	Gray	Silty cla little
-190.3-193.3	2.7	19	D-3	104	Gray	Silty cla
-197.8-200.8	2.3	20	D-3 E-2	110 -	Gray Gray	Silty cla Fine silt
-205.5-208.5	2.3	21	E-2	119	Gray	} Silty sa
-213.3-216.3	2.3	22	E-2	118	Gray	
-221.1-224.1	2.7	23	E-2	116	Gray	Silty clay
-229.1-232.1	3.0	24	E-3-a	131[x]	Grn-gray	Silty san
-237.2-240.2	3.0	25	E-3-a	133	Grn-gray	} Stiff fi
-245.3-248.3	2.3	26	E-3-a	129	Grn-gray	
-253.5-256.5	2.0	27	E-3-b	130[x]	Gray	Coarse r
-261.7-264.7	2.0	28	E-3-b	128	Gray	Medium gravel
-270.1-273.1	2.3	29	E-3-c E-3-c	- 114	Gray Gray	Clay; or Silty cla
-278.3-281.3	2.0	30	E-3-c	120	Gray	} Silty cl
-286.8-289.8	2.3	31	E-3-c	99 to 111	Gray	
-294.0	-	31C	E-3-c	-	Gray	
-296.3-299.3	2.3	32	E-4	130	Grn-gray	} Fine sil
-305.6-308.6	2.3	33	E-4	125	Grn-gray	
-313.8-315.3	1.0	34	Jfs	152	Gray	} Top of Medium
-319.7-320.2	0.7	35	Jfs	-	Gray	

Table 2 for explanation.)

Remarks
Mud (silty clay & round medium sand grains)
Mud (silty clay & medium sand)
Medium silty sand, little clay; round grains
Sandy clay; organic matter Clay
Silty clay with thin sand lenses; organic mat. Clay, little sand; peat
Firm sandy clay, many round coarse sd. grains Clay, little sand
Silty clay; shells
Silty clay, little sand — few shells
Silty clay, little sand — little organic matter
Silty clay, little sand — little organic matter; few shells
Silty clay, little sand — organic matter
Silty clay & thin fine sand lenses
Silty clay, little fine sand; shells; Foraminifera
Silty clay, thin fine sand layers; shells
Silty clay, few fine sand layers; few shells
Sandy clay
Clay, little sand — few shells
Clay, little sand
Silty clay with thin fine sand layers; little organic matter
Silty clay, few thin fine sand layers
Silty clay, little sand; few shells (1 unit wt. of 119*)
Fine silty sand & clay
Silty sand & fine clay — little organic matter
Silty sand & fine clay
Silty clay; thin fine sand lenses; tr. org. matter
Silty sand, little clay (medium fine)
Stiff fine sandy clay
Coarse round sand, little clay
Medium fine silty sand; few round grains; gravel indicated in log (1 unit wt. of 147*)
Clay; organic matter Silty clay; plants
Silty clay, little sand — few plants
Silty clay, little sand — plants
Silty clay, little sand — plants
Fine silty sand & clay
Top of rock - 314.0 Medium grained sandstone; unweathered

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weight of individual layers within the formation varies, as is shown in the logs (Tables 2, 3, 4).

Texture.—The texture of each sample was determined in the field, and by microscopical examination in the office. Mechanical analyses on representative samples were made with sieve and hydrometer according to ASTM standards Designation D422-39 (1946, p. 650–661). The various sediment parameters, particularly the median diameter (D_{50}), the coefficient of sorting, the logarithm of the skewness, and the uniformity coefficient of the 696 mechanical analyses, have been calculated from the grading curves (Table 6) and are on file in the Division of San Francisco Bay Toll Crossings. For definitions of the median diameter and other parameters, see Krumbein and Pettijohn (1938, p. 212–267) and Terzaghi and Peck (1948, p. 21, 22).

The median diameter (D_{50}) represents the mid-point on the grading curve; half of the weight of the sediment is composed of particles having diameters greater than the median, and half is smaller. The median thus is a quantitative measure of the average texture or grain size of the sediments.

The upper part of the Bay mud is materially coarser than the lower part (Table 6); the average median diameter is 6.4 microns for member A-1, compared with 3.5 microns for member A-3. Along the Southern Crossing, member A-1 coarsens from west to east. The median diameter averages 4.8 microns in the west part and 10.5 microns in the east part.

The texture of the Merritt sand is remarkably uniform laterally. The average median diameter is 163 microns, and, for half the samples analyzed, the median diameter ranges between 125 and 205 microns. As member B-1 contains more clay than member B-2, the median diameter is smaller.

The upper part of the Posey formation, member C-1, is a silty clay having an average diameter of 5 microns, compared with 136 microns for the lower part of the formation, member C-2.

The San Antonio formation varies in texture, both vertically and laterally. The average grain size for the entire formation is approximately 5

TABLE 5.—UNIT WEIGHT OF SEDIMENTS, SAN FRANCISCO BAY

Data represent average (median) of all sample tubes for the respective sedimentary units

Formation	Member	Southern Crossing			Parallel Crossing		Estuary	All areas	Number of analyses
		West	Middle	East	West	East			
		Pounds per cubic foot							
Bay mud	A-1	94±1	102±3	102±2	93±6	96±4	96±3	98±4	448
	A-2	98±2	107±4	107±8	113±5	—	97±6	100±6	358
	A-3	95±1	100±4	98±3	99±1	—	101±2	100±3	355
Merritt sand	B-1	—	—	131±2	128±2	—	133±1	131±2	55
	B-2	129±2	127±2	131±2	126±2	132±2	132±2	131±2	323
Posey	C-1	125±5	—	124±2	111±3	—	126±2	125±3	136
	C-2	134±1	—	131±3	133±2	131±2	128±4	130±3	105
San Antonio	D-1	108±2	118±7	106±2	—	104±2	114±4	108±4	221
	D-2	121±5	116±4	109±2	117±3	107±3	114±5	111±5	232
	D-3	112±4*	112±4*	113±6	114±4	108±3	111±3	112±4	689
	D-3-b	122±2	122±5	—	—	—	—	122±4	102
Alameda	E-1	123±3	120±3	122±5	124±2	126±9	131±2	126±5	374
	E-2	126±2	127±4	124±4	137±2	117±2	130±8	126±6	224
	E-3	—	127±4	128±3	—	129±5	129±5	129±5	259
	E-4	133±2	129±3	—	—	128±3	—	130±3	79
	E-5	126±2	126±2	—	—	—	—	126±2	76

* Does not include data from sand unit D-3-b.

microns. The lower part, member D-3, is the most fine-grained, having an average diameter of 4.4 microns. This member is particularly fine-grained east of Yerba Buena Island, where the median is 2.4 microns. The San Antonio formation is relatively soft in this area and has a comparatively high water content.

TABLE 6.—TEXTURE OF SEDIMENTS, SAN FRANCISCO BAY

Formation	Member	Southern Crossing			Parallel Crossing		Estuary	All areas			Number of analyses
		West	Middle	East	West	East		Q_1	Median	Q_3	
		Median diameter, D_{50} in microns									
Bay mud	A-1	4.8	6.3	10.5	13.5	6.7	5.8	4.8	6.4	10.2	74
	A-2	7.0	7.2	7.2	154	—	—	5.1	7.2	12.2	47
	A-3	1.4	4.3	2.5	3.7	—	3.6	2.4	3.5	5.8	40
Merritt sand*	B-1	—	—	148	45	—	—	76	138	183	9
	B-2	165	182	171	91	155	167	130	166	210	46
Posey	C-1	114	—	2.8	16.0	2.8	6.7	2.9	5.0	10.2	26
	C-2	—	360	45	45	—	141	23	136	200	17
San Antonio†	D-1	2.8	11.4	2.8	—	4.5	3.4	2.4	4.4	8.0	28
	D-2	5.6	181	3.0	8.0	6.7	13.5	3.2	6.4	26	32
	D-3	6.9	5.5	3.6	5.6	2.4	4.1	2.8	4.4	7.5	93
	D-3b	81	61	‡	—	—	—	7.1	72	140	24
Alameda	E-1	12.1	8.0	11.5	5.6	11.5	76	6.7	12.4	57	53
	E-2	10.5	228	16	512	5.6	76	39	85	14.5	39
	E-3	—	8.0	11.5	—	44	9.6	4.0	12.8	46	40
	E-4	—	181	—	—	45	—	42	128	395	10
	E-5	8	2.8	—	—	—	—	3.4	5.6	9.5	9

* Includes the entire Merritt sand where member B-1 cannot be recognized or is not present.

† Does not include analyses from sand unit D-3-b.

‡ As unit D-3-b on the east side of the Southern Crossing is thin and consists mainly of clay, in this area, it has been included in the general average for Zone D-3.

The Alameda formation is variable in texture and in general is sandier than the other formations. The average grain size for the formation is 15.7 microns. The first quartile is 7.5 microns, and the third quartile, 113 microns. Thus, the spread in median diameter is large.

Atterberg limits.—The Atterberg limits are measures of the inherent plasticity of sediments. Their characteristics and significance are well described by Terzaghi and Peck (1948, p. 30–36) and Casagrande (1948, p. 921). The Atterberg tests were made according to ASTM standards, designations D423–39 and 424–39 (1946, p. 662–666). The plastic limit refers to the water content at which the sediment just begins to have plastic properties; and the liquid limit, to the water content at which the sediment ceases to be plastic and just begins to flow. The difference between the two limits is called the plasticity index. Casagrande (1948) presents data which show that the liquid limit and the plasticity index vary uniformly with respect to each other in any individual environment of deposition, but the relationships for sediments laid down under different conditions, though consistent within themselves, differ from one environment of deposition to another. The difference between individual environments may be caused by differences in clay minerals. In the sediments of San Francisco Bay, the Atterberg limits vary inversely with grain size (Fig. 7). The Atterberg limits of the sediments in San Francisco Bay plot slightly above the A-line of Casagrande (1948). This position is similar to the glacial clays Casagrande studied.

The average median liquid limit for the different formational zones is given in Table 7. The extremes for any sedimentary unit are within 10 units of the median for that unit in the table. The Atterberg limits are essentially

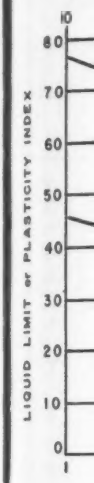

FIGURE 7.

constant laterally in the individual formations, particularly when variation due to grain size is considered. The median liquid limit of the upper part of the Bay mud (member A-1) is 56 and, for the lower part, 68. The Merritt sand and the Posey and Alameda formations have lower liquid limits, presumably because of the high sand content. The lower part of the San Antonio formation (Zone D-3) is fine-grained and has a median liquid limit of 60.

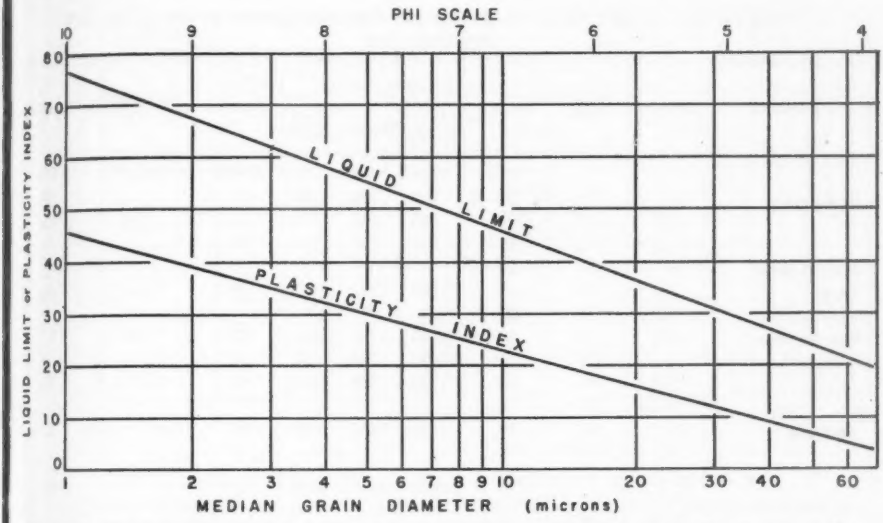

FIGURE 7.—RELATION BETWEEN ATTERBERG LIMITS AND GRAIN SIZE OF SEDIMENTS, SAN FRANCISCO BAY
Phi scale (Krumbein and Pettijohn, 1938, p. 88–90) is given at top of figure.

Shear strength.—In the present investigation, the strength of the sediments was measured by quick shear, by consolidated quick shear, and by unconfined compression. (For discussion of these tests *see* Terzaghi and Peck, 1948, p. 78–100.) The unconfined compression test is simplest to perform and when supplemented by an adequate number of quick shear and triaxial shear tests is very practicable. The strength of the sediments in San Francisco Bay as determined by quick-shear tests proves to be half the unconfined compression, which is the same relationship reported by Terzaghi and Peck (1948, p. 93).

The direct quick-shear tests were made by the standard procedure of the Materials and Testing Laboratory of the California Division of Highways, in which 2-inch cylinders of sediments were sheared in an MIT shear box, with increments of shear of .05 inch per minute. Porous drain plates were provided, and the samples were sheared both parallel and normal to bedding.

Fifteen samples were sheared by a method of consolidated quick shear in which the sediments were first consolidated under given loads and then subjected to quick shear. The samples had ample opportunity for drainage of water during the process of consolidation. The average angle *phi*, the so-called angle of internal friction, in these tests is 16°. Seven samples of the Bay mud and San Antonio formation from the west part of the bay were tested by triaxial shear. The average (median) angle of apparent internal friction was 16°.

The unconfined compression test was performed by the method of the Materials and Testing laboratory, in which cylindrical samples 2 inches in diameter and 4 inches in length are compressed between parallel plates under constantly increasing load. Stress-strain curves are plotted, and the unconfined compression determined from the point of failure (Table 7). The data for sands are not included, as uncon-

fined sand in general has little compressive strength. The results are presented in terms of pounds per square foot for half the unconfined compression ($Qu/2$).

solidation test consists of placing a cylindrical prism of undisturbed sediment in a tightly fitting cylinder and measuring the amount of consolidation for given loads for given periods

TABLE 7.—LIQUID LIMIT, SHEAR STRENGTH, AND PRECONSOLIDATION OF SEDIMENTS, SAN FRANCISCO BAY

Formation	Member	Liquid limit	Samples	Shear Strength			Samples	Preconsolidation†	Samples
				First quartile	Median	Third quartile			
			Number	*Pounds per square foot*			*Number*	*Tons/Ft.²*	*Number*
Bay mud	A-1	56	69	110	200	280	62		32
	A-2	47	32	230	300	470	34	0–0.6	25
	A-3	68	35	280	400	540	28		26
Merritt sand	B	31	10		*			*	—
Posey	C-1	49	20	1180	1800	2350	23	2.1–3.4	10
	C-2	34	10		*		—	*	
San Antonio	D-1	68	22	1050	1350	1560	27	1.4–4.6	22
	D-2	55	22	1030	1220	1650	23	1.4–4.3	9
	D-3	60	80	1230	1560	2000	88	0.3–3.1	54
	D-3-b	52	8		*			*	
Alameda	E-1	43	30	1600	2400	2900	45	−0.7–2.1‡	22
	E-2	30	8						
	E-3	43	29	2000	3000	4000		§	23
	E-4	*	—						
	E-5	55	6						

* Formation is sandy; Shear strength and preconsolidation were not determined.

† Figures give general range of preconsolidation (middle two-thirds of the size distribution).

‡ Negative figure indicates extent to which samples are underconsolidated; that is, samples have not yet been consolidated in response to imposed load. Figure 2.1 is positive.

§ Data are too variable to record.

In general, the shear strength ($Qu/2$) of the Bay mud ranges from 200 to 500 pounds per square foot. In the lower formations it increases more or less regularly with depth reaching a maximum of more than 2 tons per square foot in the Alameda formation. This increase in strength reflects normal compaction of the sediments due to overlying load. In the soft Bay mud the shear strength ($Qu/2$) increases 10 pounds per square foot per foot of depth, which corresponds to an angle of apparent internal friction of approximately 15°, and thus accords with the results from the consolidated quick shear and triaxial shear tests.

Consolidation.—The consolidation test is a measure of the capacity of the sediment to compact. It is used to predict the settlement caused by imposition of load. (For details *see* Terzaghi and Peck, 1948, p. 56–67.) The consolidation test consists of placing a cylindrical prism of undisturbed sediment in a tightly fitting cylinder and measuring the amount of consolidation for given loads for given periods of time. For any given load the sediment is allowed to reach its maximum consolidation before additional loads are imposed. Free escape of water is provided. Curves both for the rate of consolidation with time and with load are obtained.

With increase in load, the sediment consolidates or compacts relatively little until a load has been reached which corresponds approximately to the weight of overburden above the sediment at the time the sample was taken. With additional load, the degree of consolidation soon becomes constant with respect to logarithmic increments of load up to a point near the ultimate compaction. The last few per cent of total possible consolidation obeys a different law and takes place at a very slow rate as the load is increased. The constant rate of increase for loads greater than the normal load

represents the virgin part of the consolidation. It is expressed in terms of the tangent of the slope angle and corresponds to the compression index. (*See* Terzaghi and Peck, 1948, p. 65.)

in Figure 8 indicate only the order of magnitude of the settlement that may be expected in the respective formations.

To illustrate the use of these curves, let us

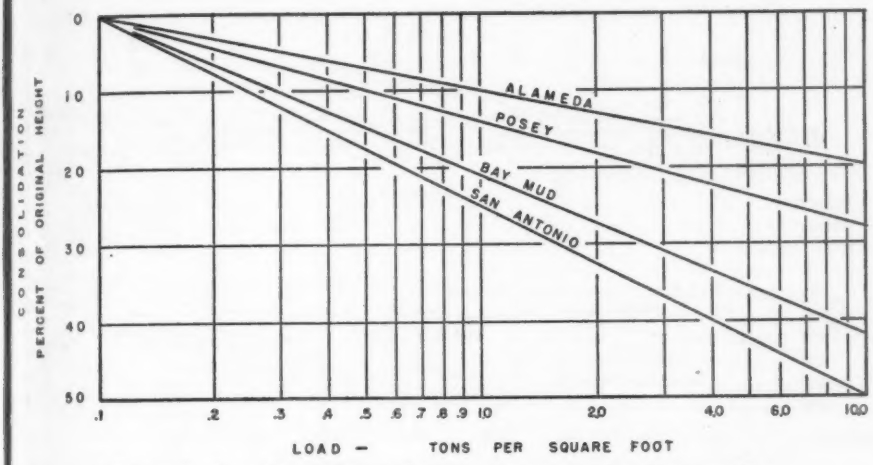

FIGURE 8.—LOAD CONSOLIDATION CURVES OF SEDIMENTS, SAN FRANCISCO BAY

The consolidation is expressed in terms of the ratio of decrease in original height of the column of sediment to logarithmic increments of load.

More than 250 consolidation tests were made. Most samples were consolidated with loads up to 8 tons per square foot, but 50 samples were consolidated up to 32 tons. The results of the samples for which three or more points lie on the virgin branch of the consolidation curve are summarized in Figure 8 which shows the average slope of the virgin branch for each of four formations. Data for the Merritt sand are not given because sand compresses relatively little.

The consolidation curves of all samples from the Bay mud have essentially the same angle of slope, and the curves of samples from the other formations do not differ greatly from the average for the particular formation. Nevertheless, since the median diameter of the samples analyzed from the San Antonio formation is 2 microns, compared with an average of 5 microns for the formation as a whole, the proper angle of slope for this formation perhaps is less than shown. The true average may be similar to that for Bay mud; that is, the slope angle of the virgin branch of the curve may be nearer .20 than .25, which latter is the slope shown in Figure 8. In any event, the curves

assume that a load of 1 ton per square foot is to be imposed upon sediments of the Bay mud at a given depth beneath the surface of the sediments, where the weight of the overlying deposits is 0.5 ton per square foot. From the curve we see that a load of 0.5 ton corresponds to a consolidation of 15 per cent. If 1 ton is added to this, the load is 1.5 tons, and from the curve we find that a load of 1.5 tons is equivalent to a consolidation of 25 per cent. The difference between 25 and 15, or 10 per cent, represents the anticipated consolidation of the Bay mud under those conditions. This figure of 10 per cent of course represents an average figure for the Bay mud. For given imposed loads the consolidation per unit thickness of sediment is greater near the surface than it is at depth, because the deeper sediments carry a greater load than the shallower deposits, and they already have consolidated more than the shallower sediments.

Some of the laboratory data on consolidation can be checked by engineering experience. As indicated in Figure 9, which shows the settlement of the fill for the mole causeway on the

east end of the San Francisco-Oakland Bay Bridge, the settlement varies almost directly with the thickness of the underlying Bay mud. The weight of the sand fill is approximately 1 ton per square foot. The average settlement, as determined by measurements taken at periodic intervals, is approximately 10 per cent of the thickness of the Bay mud; it is 8 feet where the mud is 80 feet thick and 3 feet where it is 30 feet. As the underlying Posey and San Antonio formations seem to have been preconsolidated more than 1.5 ton per square foot, they presumably have not consolidated appreciably under the added load. Skempton (1944) has shown that the consolidation varies with the Atterberg limits. A similar relationship holds for the sediments of San Francisco Bay.

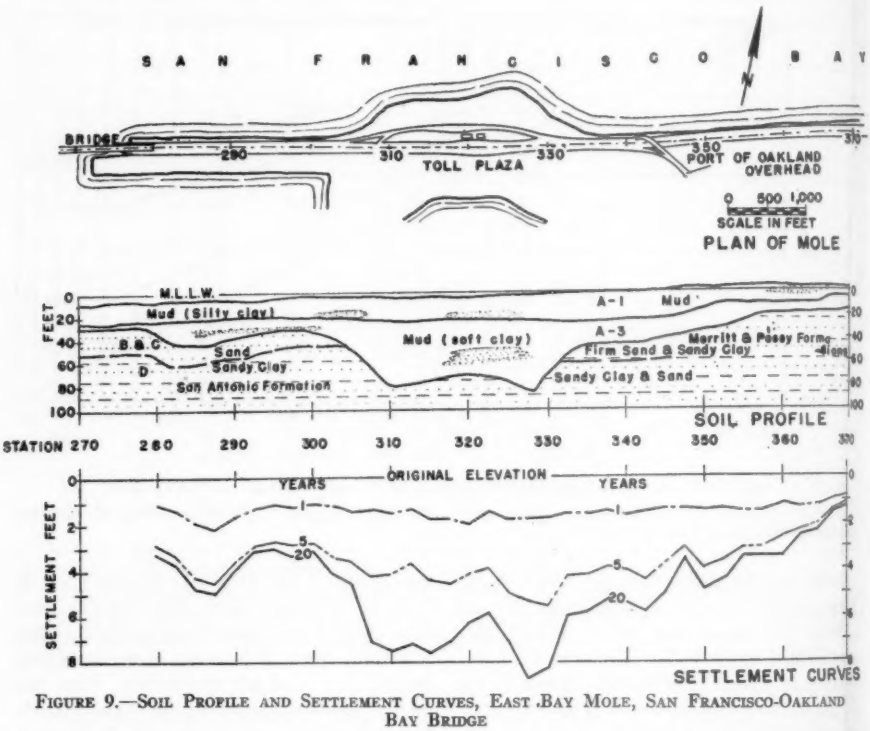

FIGURE 9.—SOIL PROFILE AND SETTLEMENT CURVES, EAST BAY MOLE, SAN FRANCISCO-OAKLAND BAY BRIDGE

Preconsolidation.— If sediments have been covered by material that has subsequently been eroded, the degree of compaction at any particular place is governed by the maximum natural load that has been added to the sediments. If some of the load is removed by erosion, the sediments expand only slightly—much less than the amount needed to occupy a volume corresponding to the normal degree of consolidation for the reduced load. Casagrande (1936) has developed a procedure for determining the extent to which the sediments have been consolidated in excess of the consolidation that would be produced by the present load. This excess consolidation is known as the preconsolidation load. The preconsolidation load is of engineering significance, because sediments ordinarily will not compress appreciably until loads greater than the preconsolidated load have been imposed. The average preconsolidation of over 200 samples is summarized in Table 7.

The Bay mud and the Alameda formation have been subject to relatively little preconsolidation, whereas the Posey and San Antonio formations have been preconsolidated 1.5 to 3

tons per square foot. If erosion caused this amount of preconsolidation, 40 to 100 feet of sediment must have been removed. Deep valleys were carved in the Posey formation prior to the deposition of the overlying sand, but seemingly little material was eroded from the areas between the valleys, as the contact between the Posey and Merritt formations in such places is nearly flat. Therefore, unless some sort of blanket erosion occurred, the preconsolidation should be explained in some way other than by removal of load. As the Posey formation seemingly was exposed to air during the subsequent period of erosion, perhaps the consolidation is some way associated with desiccation.

TABLE 8.—SUMMARY OF LABORATORY TESTS OF ROCKS IN FOUNDATIONS FOR PARALLEL CROSSING, SAN FRANCISCO BAY*

(Analyses made by Materials and Testing Laboratory, Calif. Division of Highways)

Hole	Location Pier	Elevation M.L.L.W.	Specific gravity	Bulk density lb/ft³	Per cent voids	Modulus of rupture,	Modulus of elasticity	Unconf. comp. p.s.i.	Bond with concrete‡ p.s.i.		Rock
									7 days	28 days	
52	W-1	−22	2.85	175	1.13	2940	3.9	7300	185	212	Dense greenstone
53	W-1	−34	2.78	172	0.58	—	4.1	8350			Dense greenstone
57	YBI†	+149	2.75	161	5.84	—	2.2	12,900			Weathered sandstone
57	YBI	+67	2.73	167	1.76	—	4.7	19,900			Sandstone
57	YBI	+23	—	165	—	—	4.8	9620			Siltstone (graywacke)
58	YBI	+6	2.86	167	6.18	3890	5.4	27,800	155	290	Sandstone
61	W-3	−229	—	167	—	—	1.9	3500			Weathered sandstone
62	W-4	−228	2.61	159	1.85	—			90§	195	Siltstone
62A	W-5	−173	—	169	—	—					Sandstone
62A	W-5	−160	—	165	—	—					Shale
63	W-6	−175	—	169	—	—					Chert

* *See* text for explanation of tests.
† Anchorage on Yerba Buena Island.
‡ Average tensile strength of mortar after 7 days is 395 p.s.i.; after 28 days, 508 p.s.i.
§ Rock fractured.

Rock tests.— A series of 11 samples from the rock-coring operations along the foundations for the suspension bridge over the west part of the bay were tested in the Materials and Testing Laboratory in Sacramento. The results of the test are presented in Table 8. The rocks selected for testing represent the principal types encountered in the foundations of the Parallel Crossing on the west side of the bay. The sandstone selected for these tests is massive well-indurated medium- to coarse-grained feldspathic sandstone. Two of the sandstone samples, from near the surface of the bedrock, are slightly weathered. The siltstone is a dense well-indurated fine-grained rock, interbedded in thin layers of sandstone. The shale contains a small amount of organic matter and shears relatively easily. The greenstone is a dense massive rock with an aphanitic or finely crystalline groundmass containing phenocrysts of plagioclase and augite. In places the rock contains pyrite. The greenstone is cut by many veins of carbonate mixed with minor amounts of quartz. Quartz veins up to half an inch thick are found in all rocks. Most of the greenstone of the Franciscan formation is only moderately strong, as it is broken by numerous fractures; but the greenstone in the cores from the proposed pier sites along the Parallel Crossing is dense and firm. The tests were made according to standard procedures of the Laboratory in Sacramento following ASTM practice. The modulus of rupture was performed on specimens of approximately 1-inch square cross section and 3.75

inches length, with center loading and a span of 3 inches. The modulus of elasticity and the unconfined compression tests were made on cylindrical specimens 2.14 inches in diameter and 3 to 4 inches high. The load was applied at the rate of 500 pounds per minute.

The tests on bond of cement mortar with rock were conducted by embedding 1-inch cubes of rock in the usual cement briquet molds, one end of the rock prism being on the waist line of the briquet. The upper half of the briquet space around the rock prism was filled with mortar in contact with the inch section of the rock. The mortar consisted of one part Golden Gate cement to two parts Perkins sand by weight. Specimens were cured 1 day in moist air and 6 and 27 days, respectively, in water.

Engineering Applications of Geology

Relation of Grain Size to Engineering Properties of Sediments

Laboratory tests indicate that the grain-size distribution influences materially the engineering properties of the sediments. The water content varies inversely with the particle size (Trask, 1932, p. 82). In San Francisco Bay the strength of the fine-grained sediments varies both with the water content and with the grain size (Trask and Rolston, 1950). The equation for the relationship is

$$S_s = \mathrm{Log}_{10}^{-1} \frac{7.5 - 3w - 4\ \mathrm{Log}_{10} D_{50}}{1.8 - \mathrm{Log}_{10} D_{50}}$$

where S_s is the shear strength expressed in pounds per square foot; w is the moisture content expressed as the ratio of weight of water to weight of solid particles; and D_{50} is the median diameter in microns.

The Atterberg limits of the sediments in San Francisco Bay vary inversely with the grain size (Fig. 7). The ignition loss of the sediments in San Francisco Bay also varies inversely with the grain size as in other sediments (Trask, 1932, p. 33). In addition, the consolidation, or compaction as it is called by geologists, depends to a considerable extent upon the particle-size distribution.

Consolidation, however, is a complicated subject. Basically, consolidation consists of decrease in volume as a result of forced removal of water and other material from the pore spaces of the sediment. Both porosity and permeability therefore are factors. Three statistical parameters of grain-size distribution—1, median diameter; 2, coefficient of sorting (uniformity coefficient or standard deviation of particle size); and 3, skewness (relative concentration of particle size on either side of the median)—consequently must be considered in any attempt to understand the fundamental nature of consolidation. (For discussion of these parameters *see* Krumbein and Monk, 1942, Krumbein and Pettijohn, 1938, p. 228–267, Russell and Dickey, 1950, Terzaghi and Peck, 1948, p. 21, 22, and Trask, 1932, p. 70–72.) Ultimately a quantitative relationship will be found between grain-size distribution and consolidation, but at present the problem must be approached somewhat empirically. The type of clay mineral also is a factor, but the effect of type of clay minerals is beyond the scope of this paper.

A sand, with its grains resting closely upon one another, obviously can consolidate very little; whereas a clay, which consists largely of flaky and tabular minerals separated by varying amounts of water, can compact very much. This feature is illustrated by the compression-index curves shown in Figure 8. The degree of compression of the Alameda formation, which is relatively rich in sand and contains less clay than the other formations, is much less than for the San Antonio formation, which has a low sand and a high clay content. The relationship is analogous to the relationship between compression index and liquid limits, described by Terzaghi and Peck (1948, p. 66). If the liquid limit varies with grain size and the compression index varies with the liquid limit, then the compression index should vary with the grain size. Figure 7 shows that the liquid limit varies inversely with grain size. Consequently, the compression index can be expected to vary similarly with respect to grain size.

The consolidation of a sedimentary unit can be estimated on the assumption that the average degree of consolidation of the entire formation or constituent parts is the same as the average consolidation of the samples from the sedimentary unit for which laboratory determinations were made. If the texture is relatively

uniform, the estimate obtained in this way is reliable, but if the texture varies, as it does in the sedimentary units in San Francisco Bay, the estimates obtained may be misleading.

the formational unit is 50 feet thick, that it consists of 100 per cent of clay, and that, for a given imposed load, the consolidation is 10 per cent, as determined by the average of

TABLE 9.—CLAY CONTENT OF SEDIMENTS, SAN FRANCISCO BAY
(In per cent)

Formation	Zone	Southern Crossing			Parallel Crossing		Estuary	General average
		West	Middle	East	West	East		
Bay mud	A-1	90	95	80	80	80	90	85
	A-2	60	45	45	45	—	30	45
	A-3	100	85	95	95	—	100	95
Merritt sand	B-1	—	—	15	40	35	60	40*
	B-2	5	0	5	15		15	10
Posey	C-1	30	—	85	65		85	65*
	C-2	—	0	5	10		10	10
San Antonio	D-1	95	95	100	90	90	80*	90
	D-2	35	40*	80	40*	75	75	55*
	D-3	85	80	85	70*	90	85	80
	D-3b	15	20	55	—	—	—	20*
	E-1	55	60*	50	70*	75	45*	55
	E-2	15	15*	55	5	50	10	15*
	E-3	—	50*	45	0	45	65	50
	E-4	0	15	—	—	50	—	15*
	E-5	80	80	—	—	—	—	80

Data indicate average clay content of sedimentary units. For this purpose sandy clay is rated as 50 per cent clay. General average in column 9 is weighted to give most representative estimate of general conditions in San Francisco Bay.

* Clay content of individual layers within the formational unit, or in different parts of the bay, deviates materially from the average given.

Consolidation ordinarily is determined on samples classified as clay. Sand layers interspersed with the clay consolidate very little or not at all. Consequently the percentage distribution of layers of sand and clay in the sedimentary unit should be known. Sandy clay consolidates less than clay but more than sand. Hence, in the absence of specific consolidation tests of sandy clay samples, some compensation for such samples should be made.

Tables 9 and 10 contribute data for estimating consolidation. These tables give the average percentage of clay and the average thickness of each of the sedimentary units in six areas in San Francisco Bay. In the compilation of this table, sandy clay layers were arbitrarily considered as consisting of 50 per cent sand and 50 per cent clay.

To illustrate the use of Tables 9 and 10 in estimating consolidation, let us assume that

laboratory tests of clay samples for that formational unit. The resulting compaction would be 10 per cent of 50 feet or 5 feet. If, according to Table 9, the formation contains 50 per cent clay, the resulting consolidation would be 50 per cent of 5 feet or 2.5 feet. This inference is based on the assumptions that the sand layers do not compact and that, if sandy clay layers are present, they will consolidate half as much as corresponding clay layers. This latter assumption is based on the arbitrary classification of sandy clay layers as 50 per cent clay and 50 per cent sand. Obviously the assumptions are not strictly correct, but calculations made in this way should be more reliable than calculations based on the assumption that the laboratory tests of consolidation give the correct average for the formation as a whole.

In the process of computing Table 9, the

content of clay for each formation was found to be remarkably consistent from one bore hole to another. As each figure in the table is based on information from four to nine bore holes, the data presented should indicate the order of magnitude of the clay content for the particular sedimentary unit. Moreover, the information in Table 9 accords with median grain size as shown in Table 6. Naturally, Table 9 should be interpreted with due respect to its method of computation. The table is intended as a semiquantitative effort to solve a complex problem.

TABLE 10.—THICKNESS OF FORMATIONAL UNITS, SAN FRANCISCO BAY

Data indicate observed range in thickness

(In feet)

Formation	Zone	Southern Crossing			Parallel Crossing		Estuary	Average thickness
		West	Middle	East	West	East		
Bay mud	A-1	25–45	15–35	10–35	10–65	10–25	0–40	30
	A-2	30–40	5–35	0–10	0–20	0	0–10	13
	A-3	0–30	15–35	0–40	0–30	0	0–45	15
Merritt sand	B-1	0	0	0–20	0–10		0–15	8
	B-2	0–20	0–18	0–60	0–10	20–30	10–30	15
Posey	C-1	0–10	0	0–30	0–10		0–15	5
	C-2	0	0	0–10	0–15		0–10	5
San Antonio	D-1	0–40	0–10	0–10	0–3	30–40	0–10	10
	D-2	0–7	0–10	10–20	0–10	30–40	0–15	12
	D-3	25–55	20–55	25–60	0–25	25–50	5–40	35
	D-3b	0–20	5–30	0–5	—	—	—	15*
Alameda	E-1	35–55	0–40	30–40	0–10	0–15	5–20	20
	E-2	15–40	10–50	40	0–20	15–25	0–10	25
	E-3	0	40–50	20+	0–20	55–70	45+	30
	E-4	30	10	—	—	20		20
	E-5	80+	25+	—	—	—		80+

* Average for west and middle parts of Southern Crossing only.

Bedrock Problems

The problems of the bedrock relate primarily to the strength of the rock for foundation purposes. The bedrock in the area under consideration consists entirely of the Franciscan formation. This formation commonly has been broken by many small faults, which are no longer active. In many places the Franciscan formation gives rise to landslides, but no serious landslide problems seem to be connected with the proposed crossings of San Francisco Bay. The main question is the strength of the rocks in the foundations for the piers, approaches, and anchorages for the suspension cables. The compressive strength of the rock encountered in bore holes at the sites of the proposed foundations is adequate to support the designed loads, provided the foundation is not located where the surface of the bedrock slopes appreciably. A surface of considerable relief was developed on the Franciscan rocks prior to the deposition of the Alameda and younger formations (Fig. 1). The placement of foundations upon steeply sloping parts of the old topography thus should be avoided. This question is of particular concern in foundations for the proposed piers for the suspension bridge west of Yerba Buena Island. These piers lie near the crest of an old ridge extending from Mission Rock to Yerba Buena Island. Bedrock contours were drawn in the vicinity of the proposed piers upon the basis of data from available drill holes. Two of these six piers—piers W-1 and W-6 (Fig. 4)—represented by holes 53 and 63, perhaps may be located near the crest of buried hills, and additional drilling is warranted. Pier W-6, just west of Yerba Buena Island, is further compli-

cated because underlying rocks contain fault gouge and serpentine. Considerable firm massive greenstone was found at the site of Pier W-1, but this rock is so strong that it is not likely to cause trouble.

As the surface of the bedrock at the base of some of the piers slopes 10 to 20 per cent, a difference of elevation of as much as 20 feet can be encountered across the bottom of the caissons. The excavation of this rock to form a level base for the caisson causes a particular problem at Pier W-2 where the bedrock is massive sandstone, which is hard to excavate. Piers W-3, W-4, and W-5 are underlain by alternating sandstone and siltstone, with occasional chert and shale, which pose no particular foundation problems.

Another problem is the relative ease of excavation of the rock beneath caissons, where soil and depth of weathering may vary appreciably.

The anchorages for the suspension cables must have firm foundations. In the present bridge, the west end is anchored in a gravity monolith resting on steeply dipping sandstone, siltstone, and chert. Similar rocks are found at the site of the west anchorage for the Parallel Crossing. The main consideration here is that the rock be strong enough to stand the downward and lateral thrusts of the concrete anchorage. At the east anchorage on Yerba Buena Island, where the rocks are steeply dipping massive sandstone, the chief requirement is that the rocks be sturdy enough to resist the pull of the cables, which are anchored firmly in the rock itself. The rocks in the anchorage for the Parallel Crossing are fully as strong as in the present crossing.

The roadway on Yerba Buena Island is carried through a tunnel in the massive sandstone encountered in the present anchorage. The main question in planning the tunnel for the Parallel Crossing is the thickness of sand above the bedrock. In places the rock is covered by 50 to 100 feet of wind-blown sand, which presumably is equivalent to the Merritt sand. This sand obscures the site of the portals and possibly might fill an old valley in the rock overlying the tunnel. Hence the thickness of the sand along the line of the tunnel needs to be determined in advance of driving the tunnel.

The problem of the foundations for the approaches to the bridge in San Francisco resolves itself into the question of depth to bedrock. Adequate foundations for approaches exist where the bedrock contours are above sea level. In such areas, once depth to bedrock has been established, the main complication is the thickness of the wind-blown sand overlying the rock. This sand offers a good foundation in all places except where the bedrock slopes steeply. One such locality is on Bryant Street south of the junction of the First Street approach ramp to the present bridge.

The general variation in depth to bedrock is indicated on Figure 1, but for detailed information the reader should consult Whitworth (1932).

In San Francisco in some places where bedrock lies materially below sea level, soft silt and clay have been deposited. The correlation of this clay is uncertain, because masses of wind-blown sand of varying thickness have been deposited upon it. As the mud does not seem to have accumulated above the present shore line, foundations where bedrock lies above sea level do not need to be supported on piles, but where the bedrock lies below sea level, explorations should be made to determine the distribution of mud and sand beneath proposed foundations.

Alameda Formation

The Alameda formation is very firm and consolidates relatively little. The caissons for Piers E-3, E-4, and E-5 east of Yerba Buena Island (Fig. 4) rest upon the upper part of the Alameda formation. Investigations of the probable settlement of these piers, made prior to their construction, indicated only a nominal settlement. Experience has amply confirmed the predictions. Piers E-4 and E-5 have settled less than 0.05 foot in the 15 years since they were constructed. Pier E-3, the main support of the east side of the cantilever bridge, has settled approximately 1 inch in this time. The fatigue load for these piers is slightly more than 2 tons per square foot.

Since the soil tests and geologic investigations indicate that the Alameda formation is essentially as strong in other places as it is beneath these piers, it would seem to follow that, if

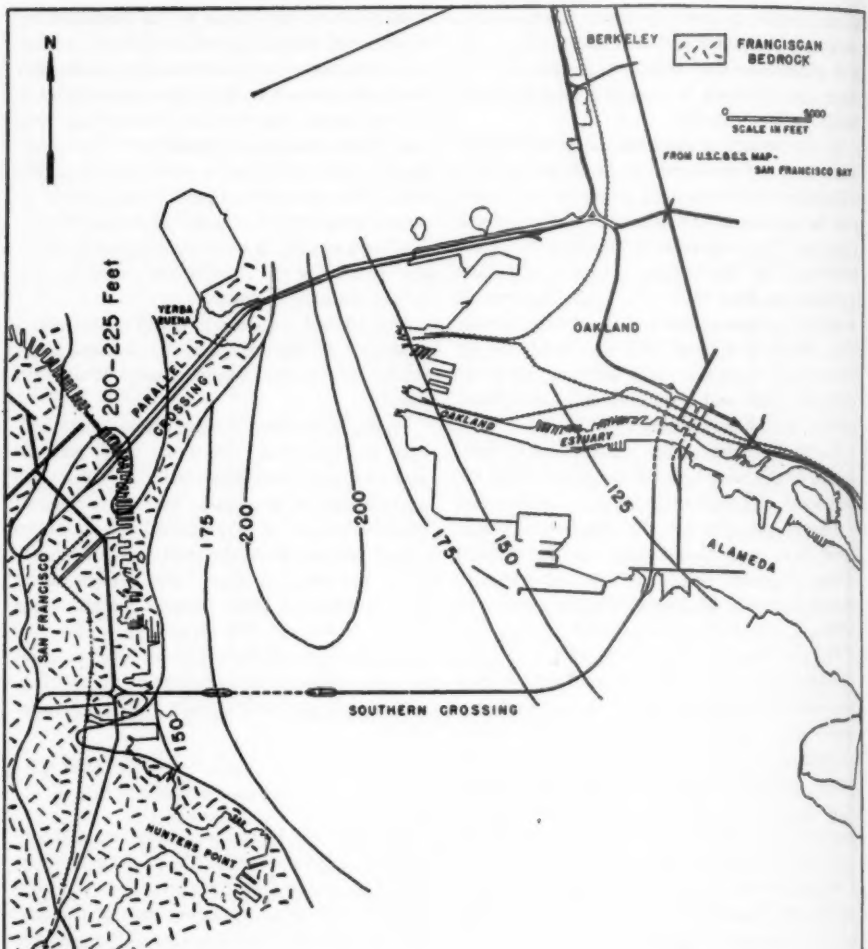

FIGURE 10.—DEPTH CONTOURS AT TOP OF ALAMEDA FORMATION, SAN FRANCISCO BAY
Alameda formation is absent over area characterized as bedrock.

design loads do not exceed those for the Bay Bridge, the formation can support the same loads. Figure 10, which shows contours at the top of the Alameda formation, accordingly gives an idea of the minimum depth at which it would be practicable to land caissons for bridges across different parts of San Francisco Bay.

Figure 10 also helps in estimating the maximum length of piles in areas where the San Antonio formation is soft, because the underlying Alameda formation is strong and will support heavy loads.

San Antonio Formation

The San Antonio formation, according to the samples tested, has been preconsolidated 1.5

to 3 tons per square foot. Consequently it should settle relatively little when fatigue loads of less than 1 ton per square foot are imposed. The San Antonio formation is softest east of Yerba Buena Island (Fig. 4). As its sediments have a relatively high coefficient of consolidation, the formation should consolidate to an appreciable extent when fatigue loads in excess of the preconsolidated load are added. Caution should therefore be exercised in dealing with this formation. For example, along the site of the Southern Crossing beneath the underwater tubes on the west side of the Bay, the San Antonio formation may consolidate as much as 2.5 feet under the load of the huge sand island that is to be imposed upon the formation (Tudor, 1948, p. 96).

The sand and gravel layers in the San Antonio formation form a good base on which to land piles. The clay layers, as a rule, give good lateral support for piles. Local variations in the properties of the sediment should however be considered when attempting to drive piles in this formation. Pile length among other things is determined by the desired load, the diameter and length of pile, and the skin friction of the formation. Skin friction can hardly be assumed to be more than the shear strength ($Qu/2$) of the adjacent sediments. Its true value presumably is some fraction of the shear strength, perhaps somewhere between $Qu/4$ and $Qu/3$.

The strength of the San Antonio formation played an important role in the decision to use pile supports for Pier E-5 in the design of the Parallel Bridge (Fig. 4). Considerable money could be saved by such a procedure. The question resolved itself (1) as to the additional stress that would be applied to the sediments underlying Pier E-5 compared with the stress under Pier E-6 of the present bridge, which had supported the design loads with comparatively little settlement; and (2) as to the ability of the sediments beneath Pier E-5 to support the additional stress. As the engineering calculations showed that this additional stress was only a relatively small proportion of the total stress beneath the existing piers, and as the sediments had ample strength to support the present piers and were of a similar character beneath the proposed pier, it was obvious that piles could be used.

Posey Formation

The Posey formation consists of two members, a firm clay or sandy clay above, and a sand or clayey sand below. The sediments have been preconsolidated 1.5 to 3 tons per square foot. Where the underlying San Antonio formation is sufficiently strong, as near the Oakland Estuary, the Posey formation affords a firm foundation. Where the overlying Merritt sand is a source of borrow sand, the Posey formation also is a potential source of sand, provided the formation is not buried too deeply or the upper clay member is not too thick. Boring operations, however, would be desirable in advance of excavation.

Merritt Sand

The Merritt sand is the main source of borrow material in the vicinity of the proposed crossings. The areal distribution of borrow sand is shown in Figure 11. This sand is relatively uniform in texture and averages 163 microns in median diameter (Tables 6 and 11). In places, particularly north of the present bridge, it contains appreciable quantities of silt and clay. West of the city of Alameda it is more than 50 feet thick. In places it is overlain by considerable thickness of Bay mud (Fig. 11), which must be removed before the sand can be dredged. Drilling in advance of dredging operations would be desirable to indicate the extent of overburden and thickness and quality of sand.

The sand commonly is thin in the valleys cutting the Posey and older formations. Little sand is found on the San Francisco side of the bay, presumably because there was no source for the sand. Most of the sand seemingly was deposited on land by the wind before it reached the bay.

In placing sand on fills, the loss of material between area of dredging and area of deposition is relatively small, owing to the comparatively small content of clay in the sand. In many areas, particularly south of Alameda, the loss is only 20 to 25 per cent. However, north of the present bridge, where the formation is more clayey, losses are likely to be greater. Table 11 presents typical mechanical analyses of the Merritt and Posey sands.

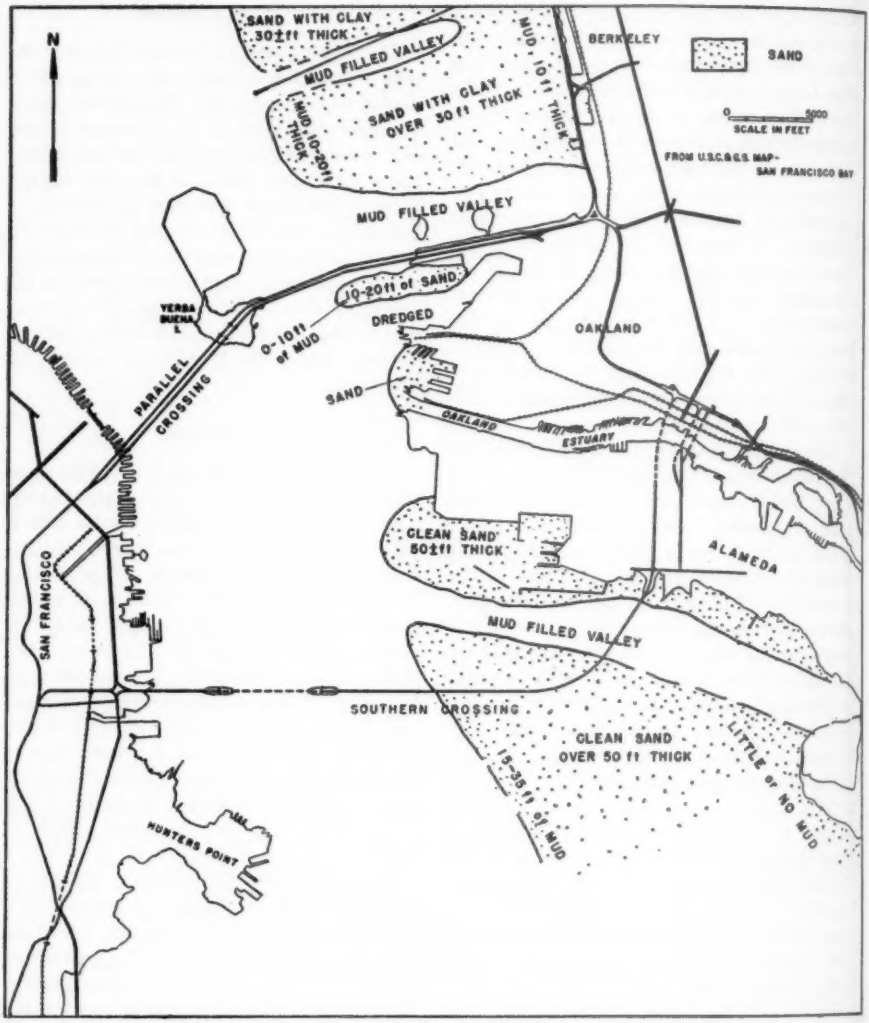

FIGURE 11.—DISTRIBUTION OF BORROW SAND, SAN FRANCISCO BAY

Thickness of mud refers to overburden above sand. Owing to lack of data, limits of sand adjacent to Treasure Island (North of Yerba Buena Island) are not shown.

Where the Merritt sand rests upon bedrock or where the underlying San Antonio formation is firm, it forms a firm foundation. In areas remote from its sources of supply, which presumably were beaches and bars exposed to the wind during periods when the sea level was lower than now, the Merritt sand contains considerable silt and in some places forms material with the characteristics of loess. One such area is near the east end of the Oakland Estuary. Here the formation when first exposed during dredging operations is firm, but within a few hours the exposed layers soften, presumably because of the disintegrating effect of the water. This feature causes complications in dredging operations for footings.

The lower part of the Merritt sand (Member B-1), in the vicinity of the tubes beneath the

TABLE 11.—TYPICAL MECHANICAL ANALYSES OF SEDIMENTS, SAN FRANCISCO BAY
(Analyses made by Materials and Testing Laboratory, Calif. Division of Highways)

Hole	Sample[1]	Depth	Formation[2]	Sand[3]	Silt	Clay	D_{50}[4]	Coef.[5] Sorting	Log Sk[6]	Crossing	Location
Number		*feet*		*Per cent*			*Microns*				
43	2-I	72	A-1	3	47	50	5.0	3.82	−0.076	Southern	West
43	9-III	123	A-3	1	32	67	2.1	3.89	.094	Southern	West
59	4-III	80	A-1	13	42	45	6.6	4.39	−0.190	Parallel	West
TP11[7]	3	36	A-1	2	50	48	5.3	—	—	Parallel	Mole
[8]	—	15	B-1	41	26	33	25	—	—	Parallel	Mole
88	3-I	68	B-1	62	3	35	160	12.4	−1.870	Parallel	Mole
47	3-IV	38	B-1	80	8	12	190	1.44	−.080	Southern	East
80	1-III	11	B-1	94	2	4	200	1.25	.000	Southern	East
485	12-II	62	C-1	4	53	43	6.7	3.42	−.075	Webster St.	South
485	13-IV	72	C-2	70	11	19	120	3.23	−.583	Webster St.	South
88	9-II	115	D-1	8	42	50	5.0	4.08	−.018	Parallel	East
88	19-II	192	D-3	6	27	67	2.5	3.33	.159	Parallel	East
484	14-I	87	D-3	24	22	54	4.1	6.40	.469	Webster St.	South
483	17-II	133	E-1	17[9]	55	27	14	3.09	−.006	Webster St.	South
42	19-II	191	E-1	36	31[9]	33	17	3.86	−1.06	Southern	West
42	21-II	219	E-2	64	26	10	81	2.45	−.436	Southern	West
44	27-VII	248	E-3	6	51	43	6.7	3.50	−.055	Southern	West
485	23-V	151	E-3	45	32	23	40	4.41	−.264	Webster St.	West
42	28-II	255	E-4	66[10]	12	6	320	3.19	.382	Southern	West
42	38-V	325	E-5	1	37	62	3.0	3.37	.074	Southern	West

[1] *See* Figures 2, 3, and 4 for location of samples. Arabic numbers represent sample numbers; Roman numbers represent particular number of the 4-inch tube within the interval of sediment sampled.

[2] *See* Table 1 for description of formation units.

[3] Sand represents material caught on 270 sieve (coarser than 55 microns); clay is material finer than 5 microns.

[4] Median diameter represents average grain size, D_{50}.

[5] Coefficient of sorting is a measure of sorting of the grains; low numbers represent well-sorted samples; that is, samples are composed largely of particles of similar grain size.

[6] Logarithm of skewness represents degree of unbalance of grain-size distribution with respect to median diameter.

[7] Bore hole TP-11 is located at the Toll Plaza (Fig. 12).

[8] Sample taken from bore hole near distribution structure, 1 mile east of Toll Plaza.

[9] Also contains 1 per cent gravel (coarser than 2 mm.)

[10] Also contains 16 per cent gravel.

Oakland Estuary, is well sorted and permeable. Consequently, water problems may be expected in cofferdams that extend down into this part of the sand.

The Merritt sand is a potential source of water. In areas of considerable industrial development, such as in the city of San Francisco, irregular draw-downs may be expected during parts of the year or from year to year. Consequently the question of subsequent exposure of piles to air, with possibility of ensuing dry-rot, confronts the engineer. Seasonal fluctuations in the rainfall also may cause similar complications.

Along the line of the Southern Crossing a thin layer of peat is found at the base of the Merritt sand. As the peat is only a few feet thick and is mixed with varying amounts of clay, it is not likely to cause serious trouble.

Bay Mud

Most of the foundation problems in the bay area relate to the Bay mud. This is the surface formation over most of San Francisco Bay. It also is the weakest formation. The foundation conditions are complicated by the variable thickness of the formation caused by the deposition of the mud in valleys (Figs. 6, 12). Near

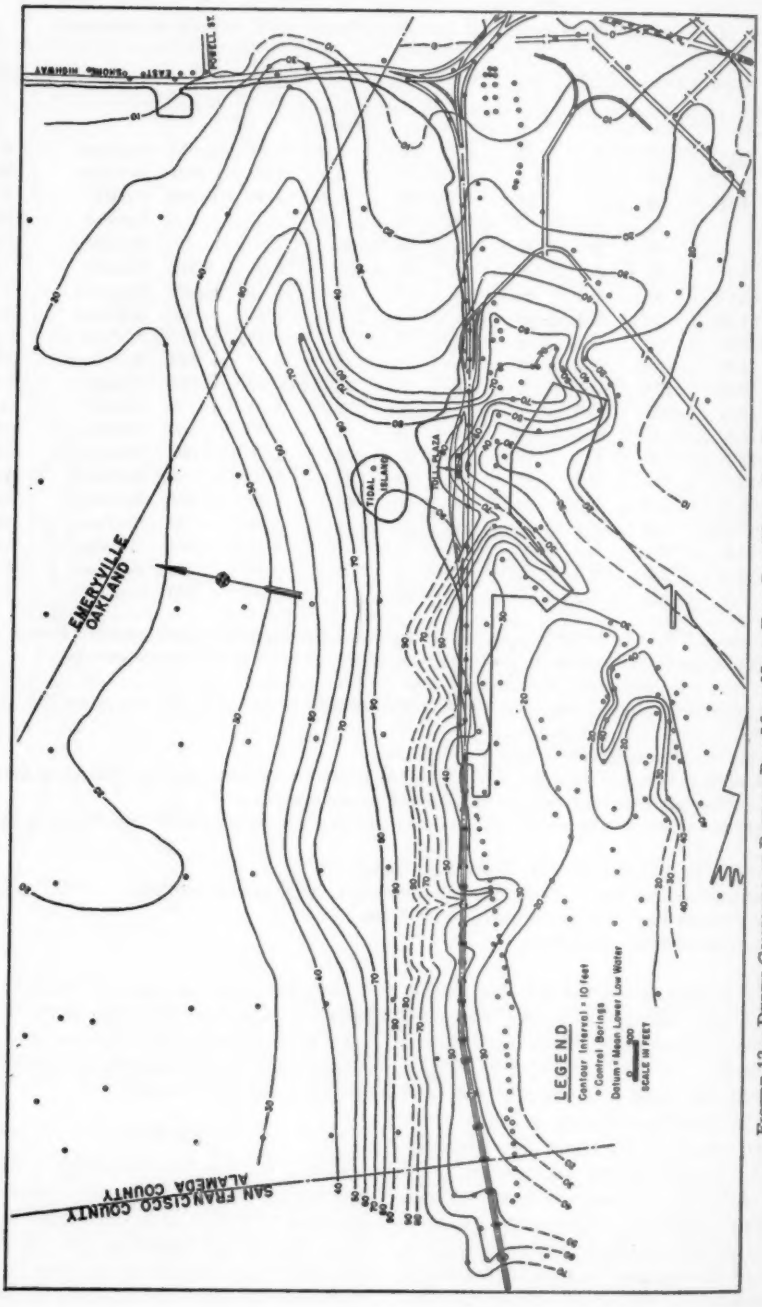

FIGURE 12.—DEPTH CONTOURS AT BASE OF BAY MUD, NEAR EAST BAY MOLE, SAN FRANCISCO-OAKLAND BAY BRIDGE

shore these valleys are filled with 80 or more feet of mud, whereas the flat areas between the valleys ordinarily are covered with less than 40 feet of mud.

The main engineering problems connected with the Bay mud are: (1) failure of the mud due to imposed loads or to excavations; (2) differential settlement; and (3) support for piles. The mud itself, according to engineering experience in the San Francisco Bay area, can support moderately deep cuts with fairly steep slopes; but when it is overloaded by fill, particularly above the water table, the mud becomes increasingly unstable as the height of fill increases. Consequently, failure ensues unless precautions are taken. In the recent excavations for the outfall sewer of the East Bay Municipal Utility District, according to a personal communication from B. B. Gordon, excavations in the Bay mud 20 to 25 feet in depth were stable with slopes of 1 to 1; in places slopes of ½ horizontal to 1 vertical were stable. Stability of deeper cuts should be calculated according to standard practice (Terzaghi and Peck, 1948, p. 182). The laboratory data indicate that a shear strength ($Qu/2$) of 250 to 350 pounds per square foot should be used in these calculations, with due consideration for the increase of strength with depth.

In the Oakland Estuary where trenches must be cut to enable the sections of the underwater tubes to be floated to their landing place, the fill now resting upon the soft Bay mud causes stability problems. Calculations indicate that benching and removal of the upper 10 feet of fill for 50 to 100 feet away from the sides of the trench would be desirable. Furthermore, material removed during excavation should not be placed on the ground between the Webster Street and the Posey tubes, because of the possibility of causing a horizontal component of motion in the underlying mud, with resulting hazard to the Posey Tube.

Unless opportunity is given for consolidation of the underlying mud, the addition of fill for mole embankments may cause failure when the height of the fill rises appreciably over 15 feet above mean lower low-water level. In the construction of the mole fill north of the Toll Plaza of the San Francisco-Oakland Bay Bridge in the winter of 1947–1948, the mud failed. In order to understand the nature of the failure, a series of holes were bored through the area of failure. Figure 13 presents a cross section through this area.

At the time of failure the sand sank at least 20 feet, and the underlying mud was forced laterally for more than 500 feet. In places the mud thrust itself upward through the previous sand fill, producing an imbricated structure. Such places of rupture are indicated by arrows in Figure 13. In this process the ground was lifted in a regular manner, but with decreasing height with respect to distance from the place of failure, much as if a hydraulic gradient had been produced in the clay.

In this failure and in other failures that occurred during the construction of the original mole for the Bay Bridge, the disturbance manifested itself for considerable distance on the down side of the plane of failure, but only for a short distance on the up side of the failure. In other words relatively little damage was incurred by sediments on the high side of the fill.

Settlement problems are largely a question of thickness of fill of the Bay mud, because, as previously discussed in connection with Figure 9, settlement due to imposed load varies roughly with the thickness of the mud. However, the mud, though in general settling uniformly, in detail does not. Consequently, differential settlement becomes a more serious problem as the thickness of the mud layer increases.

In the Webster Street crossing, the variations in thickness of the mud cause problems of settlement. The ventilation building on the south side of the Estuary in Alameda is located on the site of an old valley that has been filled with Merritt sand and Bay mud (Fig. 2). The mud is more than 80 feet thick at the ventilation building but thins rapidly toward the Estuary to approximately 25 feet. As the underlying formations are firm and the Bay mud is soft, the tube, where it rests on the mud, must be supported by piles, or in some other way to control settlement. In the Posey Tube, in places where the tube rests on appreciable thicknesses of mud, settlement has been controlled by placing the tube on a cradle supported by piles. The resulting settlement, though measurable, has been nominal.

The sections of the tube beneath the water in the Estuary have risen progressively during the 20 or more years since it was constructed.

Where the Bay mud is thick, trouble is encountered with differential settling of piles, caused by negative skin friction. The upper part

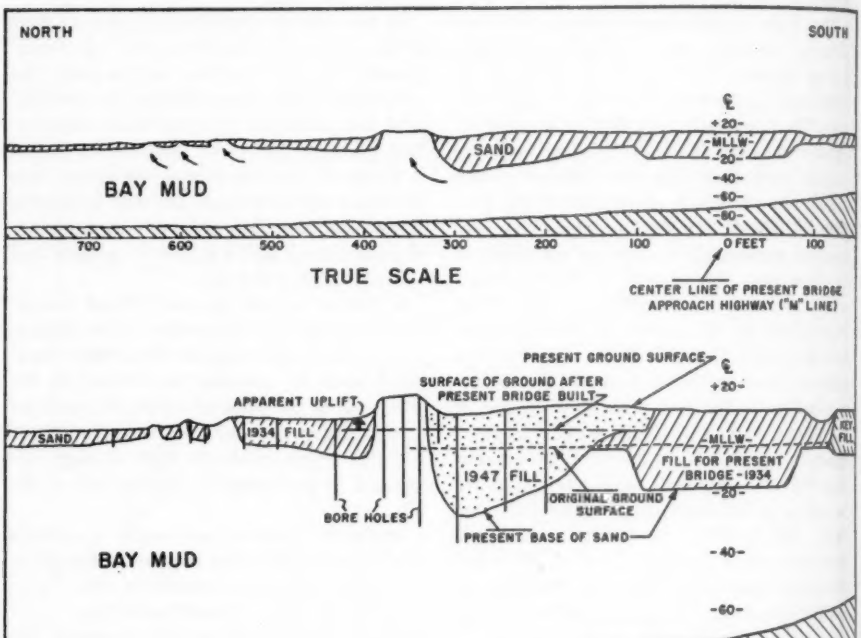

FIGURE 13.—PROFILE OF SEDIMENTS NEAR PLACE OF FAILURE, TOLL PLAZA, SAN FRANCISCO-OAKLAND BAY BRIDGE

Scale in feet.

The total rise in this time is 0.05 foot. The process in some ways seems analogous to isostasy, because the tube has risen where the overlying load is least.

The tide also seems to affect the Posey Tube, even though the tube is embedded in sand and mud beneath the water. According to F. B. Carleton, Superintendent in charge of the tube, who has made an elaborate series of tests and measurements, the tube rises and falls in direct proportion to the tide in a ratio of about one part in 600 or 700 for the change in tide, and with a lag of 30 to 40 minutes in phase.

of the mud settles more than the lower part, thus tending to drag piles downward. In many places piles have settled more rapidly than the structures they support. The prediction of the settlement of individual piles in given areas of the Bay mud is difficult. Owing to the weakness of the mud, skin friction is relatively low; consequently, as a general rule, engineers attempt to land piles in firm beds below the base of the Bay mud.

Another problem connected with the Bay mud is the rate of settlement when loads are added. The rate of settlement is influenced by

the ease with which water can escape from the sediment as it consolidates. The presence of permeable sand layers facilitates the migration of water. Consequently, where the Bay mud contains layers of sand or where sand is interspersed between laminae of clay, particularly in areas where wind-blown sand was deposited contemporaneously with the mud, settlement will be relatively rapid compared with areas where the mud contains no sand. Since the upper part of the Bay mud is sandier than the lower layer, it should settle more rapidly than the lower layer, other things being equal. The presence of sand layers should facilitate settlement when sand drains are used.

Conclusion

The present study of San Francisco Bay has indicated that the individual formations or sequences of sediments have engineering characteristics of their own, which either are essentially constant or vary with respect to the environmental conditions of deposition and the subsequent geologic history. The task of the geologist is to recognize the properties of the sediments or rocks that give them their particular engineering characteristics, and then to determine how these characteristics vary both areally and laterally within the areas under consideration. Geologists have much to learn in this respect, but the work in San Francisco Bay has thrown light on some of these questions.

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Manuscript Received by the Secretary of the Society, October 23, 1950.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 1111-1148 4 FIGS. SEPTEMBER 1951

GEOLOGIC HISTORY OF SEA WATER

An Attempt to State the Problem

(Address of Retiring President of The Geological Society of America)

By William W. Rubey

Abstract

Paleontology and biochemistry together may yield fairly definite information, eventually, about the paleochemistry of sea water and atmosphere. Several less conclusive lines of evidence now available suggest that the composition of both sea water and atmosphere may have varied somewhat during the past; but the geologic record indicates that these variations have probably been within relatively narrow limits. A primary problem is how conditions could have remained so nearly constant for so long.

It is clear, even from inadequate data on the quantities and compositions of ancient sediments, that the more volatile materials—H_2O, CO_2, Cl, N, and S— are much too abundant in the present atmosphere, hydrosphere, and biosphere and in ancient sediments to be explained, like the commoner rock-forming oxides, as the products of rock weathering alone. If the earth were once entirely gaseous or molten, these "excess" volatiles may be residual from a primitive atmosphere. But if so, certain corollaries should follow about the quantity of water dissolved in the molten earth and the expected chemical effects of a highly acid, primitive ocean. These corollaries appear to be contradicted by the geologic record, and doubt is therefore cast on this hypothesis of a dense primitive atmosphere. It seems more probable that only a small fraction of the total "excess" volatiles was ever present at one time in the early atmosphere and ocean.

Carbon plays a significant part in the chemistry of sea water and in the realm of living matter. The amount now buried as carbonates and organic carbon in sedimentary rocks is about 600 times as great as that in today's atmosphere, hydrosphere, and biosphere. If only 1/100 of this buried carbon were suddenly added to the present atmosphere and ocean, many species of marine organisms would probably be exterminated. Furthermore, unless CO_2 is being added continuously to the atmosphere-ocean system from some source other than rock weathering, the present rate of its subtraction by sedimentation would, in only a few million years, cause brucite to take the place of calcite as a common marine sediment. Apparently, the geologic record shows no evidence of such simultaneous extinctions of many species nor such deposits of brucite. Evidently the amount of CO_2 in the atmosphere and ocean has remained relatively constant throughout much of the geologic past. This calls for some source of gradual and continuous supply, over and above that from rock weathering and from the metamorphism of older sedimentary rocks.

A clue to this source is afforded by the relative amounts of the different "excess" volatiles. These are similar to the relative amounts of the same materials in gases escaping from volcanoes, fumaroles, and hot springs and in gases occluded in igneous rocks. Conceivably, therefore, the hydrosphere and atmosphere may have come almost entirely from such plutonic gases. During the crystallization of magmas, volatiles such as H_2O and CO_2 accumulate in the remaining melt and are largely expelled as part of the final fractions. Volcanic eruptions and lava flows have brought volatiles to the earth's surface throughout the geologic past; but intrusive rocks are probably a much more adequate source of the constituents of the atmosphere and hydrosphere. Judged by the thermal springs of the United States, hot springs (carrying only 1 per cent or less of juvenile matter) may be the principal channels by which the "excess" volatiles have escaped from cooling magmas below.

This mechanism fails to account for a continuous supply of volatiles unless it also provides for a continuous generation of new, volatile-rich magmas. Possibly such local magmas form by a continuous process of selective fusion of subcrustal rocks, to a depth of several hundred kilometers below the more mobile areas of the crust. This would imply that the volume of the ocean has grown with time. On this point, geologic evidence permits differences of interpretation; the record admittedly does not prove, but it seems consistent with, an increasing growth of the continental masses and a progressive sinking of oceanic basins. Perhaps something like the following mechanism could account for a continuous escape of volatiles to

the earth's surface and a relatively uniform composition of sea water through much of geologic time: (1) selective fusion of lower-melting fractions from deep-seated, nearly anhydrous rocks beneath the unstable continental margins and geosynclines; (2) rise of these selected fractions (as granitic and hydrous magmas) and their slow crystallization nearer the surface; (3) essentially continuous isostatic readjustment between the differentiating continental masses and adjacent ocean basins; and (4) renewed erosion and sedimentation, with resulting instability of continental margins and mountainous areas and a new round of selective fusion below.

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Introduction

I trust that all of you recognize that the title of this paper is largely a figure of speech. It would be interesting and even, for some inquiries, useful if we knew the history of the earth's sea water and atmosphere. But that history cannot be told until we have solved nearly all other problems of earth history. My title might much better have been "The problem of the source of sea water and its bearing on practically everything else".

Even with this modification, you still may wonder what qualifications I must think I possess to justify undertaking a subject of such global dimensions. I had originally intended to explain that I am neither an oceanographer, a geochemist, nor a Precambrian geologist, and therefore that I have no special qualifications whatever for undertaking this problem. But after discussions with various colleagues—paleontologists, petrologists, geophysicists, structural geologists, and others—I feel somewhat less modest. It seems that the subject I have selected is one in which all geologists, equally, are experts.

My interest in this general problem grew from a paper on which I began working a number of years ago. Trying to test the possibility that the phosphate rock of western Wyoming may have been laid down by direct chemical or biochemical precipitation from sea water, I began searching for some basis on which to estimate the composition of sea water and the carbon dioxide content of the atmosphere in Permian time.

It soon became evident that this question ramifies almost endlessly into nearly every fundamental problem of earth history and far beyond into the foggy borderlands between

other scientific disciplines. Eventually I began to realize some of the broader implications of the problem I had tackled. I wish to take this opportunity to review what I think I have learned about the probable history of sea water and to indicate what I think this all means in terms of general earth history.

The principal conclusions of this paper were presented orally, in December 1948 and April 1949, before the Geological Society of Washington under the titles, *The problem of changes in composition of sea water and atmosphere during the geologic past* and *A possible mechanism for the continuous supply of volatiles at the earth's surface.* Since then the main thesis has been given before various groups, in successively revised form and with different points of emphasis.

In preparing this paper I have had the unsuspecting collaboration of nearly every geologist and countless others I have talked with during the past few years. Many of my colleagues of the U. S. Geological Survey and the staff members of the Geophysical Laboratory have been most generous with their criticisms and suggestions. K. J. Murata, James Gilluly, W. T. Pecora, and D. T. Griggs have been especially helpful in leading me across some of my worst gaps of data, logic, and understanding. It is only fair to add, however, that none of these many collaborators are to be held in any way responsible for the pattern of conclusions I have tried to weave from the varied threads of our discussions.

Composition of Sea Water and Atmosphere in the Past

Lines of Possible Evidence

No more than a brief review can be attempted here of the several lines of evidence that afford at least some information about the probable composition of sea water and atmosphere in the past. I cannot refrain, however, from emphasizing the one that seems to me most promising—and most neglected. The intimate dependence of living organisms on the chemical and physical conditions of their environment is so familiar—for example, an adequate supply of oxygen in the air we breathe—that we are inclined to take it all for granted. Were it not for the small amount of ozone in the upper atmosphere, which absorbs most of the deadly ultraviolet radiations, land-living organisms could not survive in direct sunlight (Poole, 1941, p. 346; Giese, 1945, p. 226, 243–246; Allee and others, 1949, p. 74). Yet, if the amount of ozone were much greater than it is, no anti-rachitic or "sunshine" vitamin, essential to the nutrition of most animals, would be produced (Stetson, 1942, p. 18–19). Likewise, many species of modern invertebrates can tolerate only narrow ranges in the salinity of their environment, largely because of the osmotic pressures that are involved (Dakin, 1935; p. 12, 16, 25–27; Rogers, 1938, p. 670; Gunter, 1947). Many organisms have very specific requirements, on the one hand, and tolerances, on the other, for the amounts of dissolved calcium, sodium, potassium, and other elements in the waters in which they live (Rogers, 1938, p. 430, 680–682). It seems not unlikely that the ancestors of some of these modern forms may have been subject to similarly rigid requirements.

Needless to say, it is hazardous to assume that ancient animals and plants had exactly the same chemical and physical requirements as their modern descendants. Conditions may have changed gradually, and the organisms may have modified their requirements by evolutionary adaptations. Yet precisely this same hazard accompanies any effort whatever to interpret the ecology of ancient plant and animal communities. The considerable measure of success that has been attained in paleoecologic interpretations by paleontologists and ecologists, working together and balancing several lines of evidence, shows that, while the problem is difficult, it is not insuperable (Twenhofel, 1936; Vaughan, 1940, p. 457; Ladd, 1944; Lowenstam, 1948, p. 104–114, 140–142; Cloud and Barnes, 1948, p. 31, 58–66).

Paleontologists have given relatively little attention to chemical factors in the environment of ancient organisms. The most noteworthy efforts I know about have been made by a few physiological chemists (Macallum, 1904, p. 561; 1926, p. 317–322, 341–348; Henderson, 1927, especially p. 38–190; Conway, 1942; 1943). Macallum believed that the blood serum and body fluids of many animals reflect closely the composition of sea water at the time when their respective ancestral

lines became established. From this hypothesis he deduced a history of the composition of sea water. However, many physiologists do not accept Macallum's main premise. Many of you are familiar with L. J. Henderson's fascinating volume, *The fitness of the environment*, which presents the thesis that life as we know it today would be impossible if the physical and chemical conditions on the earth were much different from those of the present. All in all, it seems likely that the most definite information about the composition of sea water and atmosphere during the past will come, eventually, from the joint efforts of biochemists, ecologists, and paleontologists.

Another possible line of evidence on the composition of ancient sea water has proved much less dependable than was at first hoped. When A. C. Lane (1908, p. 63, 125) defined *connate waters*, it was with the thought that they represent samples of the original water in which a sediment was deposited. As information about these brines has grown, however, it has become increasingly evident that many chemical and physical processes have been modifying these waters since deposition of the enclosing sediments: adsorption, base exchange, dolomitization, evaporation, sulphate reduction, hydration and recrystallization of clay minerals, and other processes (Mills and Wells, 1919, p. 67–68; Newcombe, 1933, p. 189–196; Piper, 1933, p. 82, 89; W. L. Russell, 1933; Crawford, 1940, p. 1221–1222, 1317–1319; Heck, Hare, and Hoskins, 1940; Foster, 1942, p. 846–851). It now seems likely that most so-called connate waters have been so altered in composition that they do not represent at all accurately the original water of deposition.

Two other lines of evidence that afford some information on the probable composition of sea water and atmosphere in the past may be mentioned briefly: (1) Spectroscopic data on the solar and stellar abundances of chemical elements (Goldschmidt, 1938, p. 99–101, 120–121; Brown, 1949b, p. 625–627) and on the composition of the atmospheres of other planets (Wildt, 1942; Kuiper, 1949, p. 309, 326) set limits to the range of permissible speculation. (2) Sedimentary rocks of certain ages appear to have distinctive chemical or mineral characteristics over wide areas. The iron formations of the Precambrian (Eskola, 1932b, p. 39–40, 54; Leith, 1934, p. 161–164), the magnesian limestones of Precambrian and early Paleozoic age (Daly, 1909, p. 163–167), the coal beds of the Carboniferous, and other examples (Rutherford, 1936, p. 1212-1214; Landergren, 1945, p. 26–28; Lane, 1945, p. 396–398) have been mentioned as possibly indicating changes in chemical conditions on the earth with lapse of time. However, either the existence of these supposed worldwide similarities or, when their existence seems clear, this particular interpretation of their significance has been questioned (Clarke, 1924, p. 579; Leith, 1934, p. 161; Pettijohn, 1943, p. 957–960; Conway, 1943, p. 174–179, 200-202; 1945, p. 593–601, 603–604; Bruce, 1945, p. 589–590, 601; Miholić, 1947, p. 719). On the whole, it appears likely that there have been some real changes in the composition of sea water and atmosphere with time but that these changes must have been relatively small.

In the interest of brevity, I may summarize by stating that several lines of evidence seem to indicate some actual changes in the composition of sea water and atmosphere with time. Yet, the more closely one examines the evidence, the more these changes appear to be merely second-order differences. Everything considered, the composition of sea water and atmosphere has varied surprisingly little, at least since early in geologic time. This is certainly no startling conclusion to bring to an audience of geologists. It might even be considered as simply one of the underlying facts of geologic history on which the doctrine of uniformitarianism is based. Yet such a relative constancy of the composition of sea water throughout much of the geologic past has far-reaching implications, and these implications are worth a more careful consideration than they have sometimes received.

Method of Geochemical Balances

The most definite information about the paleochemistry of sea water and atmosphere may come eventually from the biologists, but the best evidence now available seems to lie in a comparison of the composition of rocks that have been weathered and of sediments that have been deposited during the geologic past. This general method has been used by Clarke (1924, p. 31–32), Goldschmidt (1933,

p. 131–133), Kuenen (1946), and others (Leith and Mead, 1915, p. 73; Twenhofel, 1929, p. 395–399) to estimate the total mass of rocks eroded and sediments deposited. Using a modification of this method, I have attempted to bring these earlier estimates up to date. But one must still rely on much the same data used by others; and attempted refinements of the method still give pretty much the same answers as before. Suffice it to say that, when this method is applied to rocks of different ages and to each of the commoner rock-forming elements—silicon, aluminum, iron, calcium, and several others—, the results are in surprisingly good agreement with one another. But for another group it is clear, even with present data, that there is no such agreement. All the constituents of this latter group are much too abundant in the present atmosphere and hydrosphere and in ancient sediments to be accounted for as simply the products of rock weathering—the explanation that fits well enough for the commoner elements. Compared with the commoner rock-forming oxides, the members of this group are all relatively volatile[1]. They are the substances which, in the language of the early chemists, we might call the *distillable spirits* of the earth's solid matter.

I had originally intended to present in this paper a detailed statement of the method by which these estimates have been made. The main paper grew out of all bounds in several directions, however, and these estimates and a fuller statement of the method by which they were derived must be reserved for publication in a separate article. The method of geochemical balances, as it might be called, can be outlined only briefly here.

In any given unit of geologic time, the quantity of material weathered and eroded from (*a*) crystalline rocks and (*b*) previously formed sedimentary rocks, plus (*j*) juvenile matter from volcanic gases, hot springs, etc., equals the quantity of material deposited as sediments on (*c*) continental platforms and (*d*) the deep-sea floor, plus (*s*) matter stored in sea water or escaped into interstellar space.

The results of applying this general equation of geochemical balances are the basis for much of the entire discussion that follows. For some elements, such as Si, Al, and Fe, which are relatively nonvolatile and sparingly soluble at moderate temperatures, the juvenile and stored-in-sea terms are probably negligible; so that for these elements the general equation can be simplified considerably. If we have reasonably good estimates of the chemical composition of crystalline rocks, older sedimentary rocks, new continental deposits, and new deep-sea sediments, we can find for each constituent the relationship between *a*, the proportion of crystalline rocks to all rocks eroded, and *c*, the proportion of continental sediments to all sediments deposited. From the percentages of two chemical constituents in each kind of rock eroded and sediment deposited, we can find one expression relating *a* to *c*. From percentages of more than two constituents in each rock and sediment, we obtain several simultaneous equations that permit a fairly rigorous test of the chemical compositions we have used.

This is all simple enough in theory, but it is difficult in practice because accurate and representative analyses of sediments are not available. It is encouraging that a National Research Council committee is now being organized to improve available information on the chemical composition of sedimentary rocks. But, even from present information, it is possible to learn something by means of this general equation (or rather the series of equations derived from it) and to obtain results that are not grossly in conflict with other types of evidence.

It turns out that, for average crystalline rocks and sediments of Mesozoic and later ages, the average chemical compositions now available agree surprisingly well with one another for most of the permanent rock-forming oxides—SiO_2, Al_2O_3, total Fe, CaO, Na_2O, K_2O, TiO_2, and P_2O_5. If the average composition of igneous rock in earlier times was somewhat nearer that of plateau basalt, this statement is also true for sedimentary rocks of Paleozoic and even Precambrian age.

This general equation gives only ratios between the several kinds of rocks eroded and sediments deposited. Three independent estimates (based on the amount of sodium now dissolved in the ocean, the quantity of continental sediments remaining uneroded today, and the mean rate of deep-sea sedimentation) accord reasonably well in their results and thus afford means for converting these ratios into quantities of rocks eroded and sediments deposited. When all is done, the final estimates have at least this to recommend them: for the greater part they fall well within the range of estimates that Clarke, Goldschmidt, Kuenen, and others have made by simpler and more direct methods.

The foregoing statements apply to the major rock-making constituents, which are the ones that agree rather well with one another in the general

[1] Hence Fenner's convenient term, "volatiles", for the entire group (1926, p. 696–697).

equation of geochemical balances. Collectively they make up from 60 to 95 per cent of all the rocks eroded and sediments deposited. But for the remaining constituents it is quite a different story. All but two of these latter form relatively volatile compounds at moderate temperatures and pressures; and it is the source of this group of volatile constituents that seems particularly significant in the geologic history of sea water and atmosphere.

TABLE 1.—ESTIMATED QUANTITIES (in units of 10^{20} grams) OF VOLATILE MATERIALS NOW AT OR NEAR THE EARTH'S SURFACE, COMPARED WITH QUANTITIES OF THESE MATERIALS THAT HAVE BEEN SUPPLIED BY THE WEATHERING OF CRYSTALLINE ROCKS

	H_2O	Total C as CO_2	Cl	N	S	H, B, Br, A, F, etc.
In present atmosphere, hydrosphere, and biosphere..	14,600	1.5	276	39	13	1.7
Buried in ancient sedimentary rocks.....	2,100	920	30	4.0	15	15
Total.....	16,700	921	306	43	28	16.7
Supplied by weathering of crystalline rocks..	130	11	5	0.6	6	3.5
"Excess" volatiles unaccounted for by rock weathering	16,600	910	300	42	22	13

Let us look at the way these "volatile spirits" are now distributed on and near the surface of the earth. In Table 1 are summaries of the best estimates I can find or make of the quantities of water, carbon dioxide, chlorine, etc., in today's atmosphere, ocean, fresh water, and organic matter. In the second row are the estimated totals of these constituents that are now buried in sedimentary rocks and that must have been part of the atmosphere and ocean in earlier times. Next we have the sums of the first and second rows, the quantities that must have come from somewhere and whose source should be accounted for. In the fourth row, we have, from the amount of crystalline rocks that must have been eroded to form all the sediments and from the composition of average igneous rocks, the amounts that must have been released by the weathering of crystalline rocks. Finally, in the last row, the differences between the quantities in the third and fourth rows, we have the "excess" volatiles that cannot be accounted for as simply the products of rock weathering.

For the sake of simplicity, C is shown as CO_2 in Table 1. This arbitrary convention oversimplifies the actual situation, for O does not occur in the right amounts to balance C as CO_2 exactly. Estimating directly (in units of 10^{20} g), the amount of O not combined in permanent rock oxides and in water, we find about 12 units in the atmosphere, 26 in the SO_4 of sea water, and small amounts in HCO_3, H_2BO_3, and elsewhere, giving a total of approximately 39 units of O in the present atmosphere, hydrosphere, and biosphere. In sedimentary rocks, estimates indicate about 490 units in the carbonates, 25 in organic matter, 21 in SO_3, 14 in the oxidation of FeO to Fe_2O_3, and 2 in the interstitial water, giving altogether about 550 units of O in the ancient sedimentary rocks. This, together with that in the atmosphere, etc., makes a total of about 589 units. Approximately 8 units of O have probably been released during the weathering of the small amount of CO_2 in crystalline rocks and 75 units from the Na, Mg, Ca, and K now dissolved in sea and interstitial water, which comes to a total of 83 units from rock weathering. This leaves (589 − 83 =) 506 units of O "unaccounted for" and is to be compared with ($\frac{12}{44} \times 910$=) 248 units of C similarly "unaccounted for".

If no other complications were involved, these quantities might be explained by the release of 483 units of CO_2 and 271 units of CO from some other unspecified source. But to find a more meaningful balance between the "excess" C and O, one must consider also the probable composition of other gases, besides CO_2 and CO, that may have come from the same unspecified source. Associated with an estimated 68 units of C and 25 units of O in the organic matter in sedimentary rocks, there should be about 9.6 units of H. If this was released originally as H_2O, it would account for 76 of the 506 units of "excess" O in the above calculation. On this assumption, the "excess" of 248 units of C and

(506 − 76 =) 430 units of O would be equivalent to 274 units of CO_2 and 404 units of CO.

It is conceivable, however, that the "excess" Cl, N, S, Br, and F shown in Table 1 were originally released as HCl, NH_3, H_2S, HBr, and HF. If so, the 19 units of H thus required have since lost their identity, presumably by oxidation to H_2O; and this oxidation would have subtracted approximately 150 units of O from the original C-O mixture. On this assumption, the original "excess" of 248 units of C and (430 + 150 =) 580 units of O would have been equivalent to 687 units of CO_2 and 141 units of CO.

In these three hypothetical mixtures of CO_2 and CO, the CO_2 ranges between 40% and 83%. If the original gases also contained significant quantities of methane, CH_4 (Poole, 1941, p. 350–351), the percentage of CO_2 would have been correspondingly higher. It is evident that the proportion of "excess" C originally released as CO_2 cannot be estimated, even approximately, unless the original composition of all the "excess" volatiles is known. For this reason, all C is shown by arbitrary convention in Table 1 as CO_2.

Possible Source of the "Excess" Volatiles

Nomenclature[2]

Only two possible sources of the "excess" volatiles occur to me: Either (a) they are largely or entirely residual from a primitive atmosphere and ocean; or (b) they have largely or entirely risen to the surface from the earth's interior during the course of geologic time.

A few years ago an eminent geologist, in discussing magmas, divided those who have views on the subject into the *pontiffs* and the *soaks*. The classification is not directly applicable to those who have views on the origin of the ocean. But here, also, we have two opposing schools of thought, and, with aplogies to Bowen (1947, p. 264), his classification may readily be modified to fit the occasion. From the very nature of the problem, there can be no anhydrous pontiffs when the origin of the ocean is considered but only soaks of one persuasion or another. Here, we may say, we have the *quick soaks* who prefer to have the wetness of the ocean there at the very beginning, all of it at once; and then there are the *slow soaks* who prefer to increase the liquor gradually by small increments over a much longer period of time. The quantities involved are the same by both courses of action, but the effects are conspicuously different: Taken in small drafts and with a proper regard for timing, an astonishingly large quantity of volatile spirits can be handled by the earth without showing it; but taken hastily, even moderate amounts are almost certain to have noteworthy effects.

[2] Literally, "name-calling".

Dense Primitive Atmosphere—Solution in the Melt

The two contrasting procedures or viewpoints just mentioned afford a basis for discussing the probable source of these "excess" volatiles. In the opinion of some who have considered the problem—those who are of the "quick-soak" school of thought—, the source of these volatiles can be deduced without too much trouble merely from a consideration of the conditions of the primitive earth. Their argument runs something like this: As the earth was probably once molten throughout, it follows that all or a large part of the water, carbon dioxide, etc., would have been volatilized in a primitive atmosphere. On subsequent cooling, the water vapor would condense into a primitive ocean; and the present hydrosphere and atmosphere are thus residual from this primitive ocean and atmosphere.

Yet, when we examine it more closely, this conclusion is not altogether convincing. To begin with, it is by no means certain that the earth was originally molten (R. T. Chamberlin, 1949, p. 252–253; Latimer, 1950; Slichter, 1950). The inert gases—neon, argon, krypton, and xenon—are from 10^6 to 10^{10} times less abundant on the earth than in the atmospheres of the stars, the lighter gases showing the greater discrepancies. Compounds and elements of the same molecular weights—water, nitrogen, carbon dioxide, carbon monoxide, and oxygen—would have been lost in the same proportions if they too had existed as gases when the earth was formed. The fact that water is present in considerable quantity on the earth today is evidence that much of it was bound in chemical compounds or occluded in solid (that is, in relatively cool) matter when the earth accumulated (Aston, 1924; Jeffreys, 1924; Russell

and Menzel, 1933, p. 999–1001; Rayleigh, 1939, p. 463; Brown, 1949a; Gibson, 1949, p. 278; Suess, 1949; Jones, 1950, p. 420, 423–424, 428–429). This leaves us with two alternatives: either the earth remained solid from the beginning, except for local melting; or it was first cold, then heated by some process until molten throughout (Urry, 1949, p. 172, 179) and finally solidified again. Whichever one of these two theories may ultimately prevail, the evidence appears distinctly unfavorable to the concept of an atmosphere that ever contained very much of the earth's total volatiles. Instead, this evidence seems to call for some mechanism by which the volatiles were largely retained within the interior of the earth, to the extent of only a few tenths of 1 per cent of the solid matter. This appears entirely possible, whether or not the earth was ever completely molten.

These interpretations are not new, but neither are they as widely accepted as I think they deserve to be. I shall try, therefore, to show why a consideration of the geologic record leads me to these conclusions.

We might start by attempting to follow through, in a semiquantitative way, some of the consequences of a dense primitive atmosphere. If all the water in today's atmosphere and hydrosphere and in ancient sediments were once present in a primitive atmosphere, and if this atmosphere were in contact with molten silicate rock, then some fraction of the water would dissolve in the melt, the amount depending on the vapor pressure, the solubility of water in the melt, and the total quantity of molten material. Using Goranson's data for the solubility of water in a melt of granitic composition, one finds that, at this vapor pressure, the melt would dissolve 2½ per cent by weight of water.

As the deeper-lying rocks probably now contain considerably less than 2½ per cent water, this must be simply a limiting case. If there were then any considerable quantity of molten material and if this were stirred by convection currents, the water that would dissolve in the melt would materially reduce that in the atmosphere until an equilibrium was reached. Making allowance for the water now in deep-lying rocks, calculation shows that a primitive atmosphere in equilibrium with a completely molten earth would contain less than one-tenth the amount of water in the present ocean. As the supposed molten earth cooled and crystallized, the water in the remaining melt and in the atmosphere would be unlikely to remain in equilibrium, for the surface would probably crust over with slag long before the interior had fully crystallized. Thus, when crystallization was completed, the primitive atmosphere, even on this hypothesis, would probably contain at that stage, as it had at the very beginning, much less water than the present ocean.

Goranson's data on the solubility of water in a melt of granitic composition at the temperatures and pressures at which crystallization begins (Goranson, 1932, p. 229–231) appear to be the most nearly applicable to the conditions here postulated. I find no data on the solubility of other gases in silicate melts of the required composition. One would expect, from the work of Morey and Fleischer (1940, p. 1051–1057) on the system CO_2-H_2O-K_2O-SiO_2, that the proportion of CO_2 to H_2O dissolved in the melt would be somewhat higher than that in the vapor phase or atmosphere in equilibrium with it, but not enough higher to modify the results of the following calculation significantly.

Under the assumed conditions, the primitive atmosphere would contain 16.6×10^{20} kg of water vapor. Over the 5.1×10^{18} cm² area of the earth's surface, this quantity of water would exert a pressure of 325 kg/cm².[3] Goranson found that a melt of granitic composition at liquidus temperatures dissolved 3% by weight of water at 400 bars[3] and 2% water at 260 bars (Goranson, 1932, p. 234). Interpolating between these values, it appears that, at a vapor pressure of 325 kg/cm², a melt of this composition would dissolve 2.5% water, and the temperature of crystallization would be about 985°C.

If the rock materials of the earth's crust retained after crystallization, and still contained today, an average of about 2.5% water, the results of this calculation would be of greater significance. Actually they now contain much less than this—only 1.15% in Clarke and Washington's average igneous rock (Clarke, 1924, p. 29). On this hypothesis, over half of the calculated 2.5% water in the melt must have escaped to the surface during or after crystallization. This means that some unspecified quantity of the water in today's atmosphere, ocean, and sedi-

[3] 1 bar = 1.01972 kg/cm² (Birch, Schairer, and Spicer, 1942, p. 319.)

mentary rocks was not there earlier and so was not present in the supposed primitive atmosphere. Hence the original vapor pressure would have been less than the 325 kg/cm² calculated from today's total of 16.6×10^{20} kg. The amount of this difference depends largely upon the quantity of melt that was involved in the assumed equilibrium.

With present data, the amount of water in the supposed primitive atmosphere can be estimated only within rather broad limits. For purposes of this calculation, the average water content of crystallized rock to a depth of about 40 km may be taken as near 1.15%. This would amount to 6.6×10^{20} kg of water. If this thickness of 40 km of rock were molten, the total water in the primitive atmosphere and melt together would be $(16.6 + 6.6 =) 23.2 \times 10^{20}$ kg. From Goranson's data it may be found that, under conditions of equilibrium, 46% of this total water would be dissolved in the melt; the remainder left in the atmosphere would exert a vapor pressure of 244 kg/cm² at the earth's surface. At this pressure the granitic melt would dissolve 1.9% water, and the melting point would be about 1010°C. If it were true that the rocks of the earth's interior now contain 1.15% water to a depth of 300 km, similar reasoning would show that about 85% of the total water would have been dissolved in the melt; the amount remaining in the atmosphere would have had a vapor pressure of 175 kg/cm², and the melt would have contained about 1.3% water.

It is extremely doubtful, however, that rocks lying more than a few tens of kilometers below surface contain as much as 1.15% water. For the rocks of the deep interior, direct determination of the water content is, of course, impossible; and there is little on which to base an estimate other than the composition of stony meteorites (Clarke, 1924, p. 42–44; Brown, 1949b, p. 627–629). Meteorites contain few if any hydrous minerals; yet an average of 63 chemical analyses, "of the highest grade obtainable", of stony meteorites gives 0.47% H_2O (Merrill, 1930, p. 16–17), and a recent digest of all earlier analyses shows an average content of 0.063% H [equivalent to 0.56% H_2O]. (Brown, 1949b, p. 626). Some of this water, in even the most carefully collected and analyzed samples, may be the result of weathering after the meteorites reached the earth. At least some, however, is probably present as original impurities or occlusions, as seems indicated by the quantities of H_2, CO_2, and other gases in both stony and iron meteorites (R. T. Chamberlin, 1908, p. 26; Farrington, 1915, p. 190–196; Nash and Baxter, 1947, p. 2541–2543). More data on the volatile content of stony meteorites are badly needed; but from present information the best estimate for the siliceous rocks of the earth's interior, from depths of 40 km to 2900 km, is probably about 0.5% H_2O.

If the iron meteorites are similarly taken to represent the materials of the earth's iron core, from 2900 km to the center, we may assume the water content there is negligible.

For rocks of the deep interior, one should logically have data on the solubility of water in melts of basaltic or peridotitic composition. However, in comparison with the uncertainties about present water content of these rocks at depth, the lack of data on the solubility of water in a more basic melt is no actual handicap. Increasing pressure at greater depth would increase very considerably the solubility of water in silicate melts (Goranson, 1931, p. 492–494). Under the assumed conditions of a molten earth and with the relatively small quantities of water involved, it is probable that nowhere would the amount of water carried down into the interior by convection currents exceed that which could be dissolved in the deeper-lying melt of different composition. The controlling factor therefore would be the solubility of water in the part of the melt that was in contact with the atmosphere.

For the densities of rocks at various depths to the center of the earth, Bullen's values (1940, p. 246; 1942, p. 28), which are calculated from seismic and astronomic data, may be used.

With these various estimates, it may be calculated that, in an earth molten to the base of the silicate mantle at 2900 km, the total water in the primitive atmosphere and silicate melt would have been $(16.6 + 207 =) 224 \times 10^{20}$ kg. Under conditions of equilibrium, approximately 98% of this total would be dissolved in the melt. The remainder left in the atmosphere would have a vapor pressure of 76 kg/cm² at the earth's surface, and the melt would contain about 0.54% by weight of water. In an earth entirely molten to the center of the iron core, approximately 99% of the total water would be dissolved in the melt. The remainder in the atmosphere would have a vapor pressure of 52 kg/cm², and the melt would contain 0.37% H_2O. This water in the primitive atmosphere, about 2.6×10^{20} kg, would be only 16% of the total in today's atmosphere and ocean and buried in ancient sedimentary rocks. If the estimate used here of 0.5% H_2O in the earth's siliceous mantle is too high, the total quantity of water involved in the equilibria would be less, and the remainder in the primitive atmosphere would be even lower than 16% of that today.

For purposes of presentation, this calculation has proceeded through steps of increasing depth of melting from the earth's surface downward to the center of the earth. Actually the hypothesis of an originally molten earth requires the reverse order,

starting with a completely molten condition and solidifying from the bottom or, rather, from the base of the silicate mantle, upward (Adams, 1924, p. 467–468). If, at any stage of cooling, the surface crusted over with slag, equilibrium between the water content of the atmosphere and melt could no longer be maintained, and the primitive atmosphere would end up with some earlier water content, which would be much less than 16.6×10^{20} kg.

If, on the other hand, the surface did not crust over before the earth solidified throughout, heat from the deep interior would continue to escape rapidly; and solidification would probably have been completed within a few tens of thousands of years (Jeffreys, 1929, p. 79, 147–148; Slichter, 1941, p. 567, 582–583). With such rapid freezing, it is improbable that equilibrium could have been maintained between the water content of the atmosphere and that of the immense volume of the melt. Thus, on this alternative also, the primitive atmosphere would probably end up with a water content significantly less than 16.6×10^{20} kg.

Dense Primitive Atmosphere—Chemical Effects

A totally different line of approach also indicates that the fraction of present hydrosphere and atmosphere that might be residual from a primitive ocean and atmosphere is probably very small. Let us assume, for purposes of calculation, that all the "excess" volatiles were once in a very hot atmosphere; that, on cooling, the water vapor condensed to liquid and dissolved all the chlorine, fluorine, bromine, etc.; and that the principal gases—carbon dioxide, nitrogen, and hydrogen sulphide—were dissolved in the water according to their solubilities and partial pressures.

This would mean at first a very acid ocean (pH 0.3). This acid water would attack bare rock with which it came in contact, and the primitive ocean would rapidly take into solution increasing quantities of bases from the rocks, until the solubility product of calcite or dolomite was reached. Thereafter, as more dissolved bases were brought in by streams, carbonate would be deposited, and carbon dioxide subtracted from the atmosphere-ocean system.

Eventually conditions would be reached under which primitive life could exist. Certain mosses have been found to tolerate as much as one atmosphere of carbon dioxide (Ewart, 1896, p. 404, 406–407), an amount more than 3000 times that in the atmosphere today. If we take this value as a basis for estimating when life and the photosynthetic production of free oxygen began, we find that sea water then would have a pH of about 7.3; a salinity approximately twice that of the present; and, because of buffer effects, considerably more bicarbonate than chlorine in solution. The bases required to balance the acid radicals, added to those previously deposited as carbonates, would call for the weathering by that time of considerably more igneous rock than appears, from other evidence, to have been weathered in all of geologic time. And, even worse, more than half the original carbon dioxide would still remain in the atmosphere and ocean.

This hypothesis also calls for the deposition of tremendous quantities of carbonate rocks before the carbon dioxide of the atmosphere was lowered to the point where life could first exist. No such excessively large quantities of carbonates are known in the earliest Precambrian sediments. Compilations of measured thicknesses of sedimentary rocks suggest that the percentage of limestone in sediments of Precambrian age is about the same as (or perhaps lower than) that in later sedimentary rocks. Sederholm's estimate (1925, p. 4) indicates a very low proportion of limestones and dolomites in the Precambrian rocks of Finland. Schuchert's (1931, p. 58, 49) figures (25% limestone in the Precambrian, 20% limestone in Cambrian to Pleistocene inclusive) are heavily influenced by estimates, now known to be excessive (Osborne, 1931, p. 27–28), of the thickness of the Grenville series in southeastern Canada and so should be lowered somewhat. Leith's estimates (1934, p. 159–162) show from 1 to 48 (average about 19%) limestone in sections of Precambrian rocks in North America, Africa, and Asia and an average of 22% limestone in sedimentary rocks of later age. There is thus no indication in the geologic record of the tremendous quantities of early Precambrian limestone that would be required under this hypothesis. These and other semi-quantitative tests make the hypothesis of a dense primitive atmosphere appear very improbable.

Under the given conditions of temperature, pressure, and composition (essentially no free hydrogen or oxygen), the quantities of the gases, CH_4, NH_3, and SO_3, that would be stable may be taken as negligible, and therefore the "excess" C, N, and S present as CO_2, N_2, and H_2S. If all the Cl, Br, and F were dissolved in 16.6×10^{20} kg water, they would amount to 0.529 acid equivalents per liter. Initially this amount of acid would be balanced almost solely by hydrogen ions, from which a hydrogen-ion concentration of $0.529 = 10^{-0.277}$ or a pH of 0.277 may be calculated.

This acid water would decompose the rocks with which it came in contact, and in so doing it would dissolve some of the bases. As a result, its acidity would decrease, and the chemical effects of dissolved gases would come to play an important part in the over-all balance. Better data are available for estimating the required coefficients at moderate temperatures than at those near the boiling point of water; and 30°C is here chosen as the basis for calculation. At this temperature and in sea water of normal salinity (35‰ or a "chlorinity" of 19.4 ‰), the solubility coefficients of CO_2, N_2, and H_2S are estimated at 0.026, 0.0009, and 0.077 mols/liter (Harvey, 1945, p. 66; Sverdrup, Johnson, and Fleming, 1946, p. 191); the first and second apparent dissociation constants of CO_2 at 1.11×10^{-6} and 1.27×10^{-9} (Harvey, 1945, p. 64); the two comparable dissociation constants of H_2S at 3.0×10^{-7} and 2.8×10^{-14} (Latimer, 1938, p. 316); the single dissociation constant of H_3BO_3 at 2.26×10^{-9} (Harvey, 1945, p. 62); and the apparent solubility product of $CaCO_3$ at 1.2×10^{-6} (Revelle and Fleming, 1934, p. 2091; Wattenberg, 1936, p. 176; Smith, 1940, p. 182). Equations that express the relationships between various quantities involved in the CO_2 equilibria of sea water (Harvey, 1945. p. 61–68) may, with some patience, be solved by the method of successive approximations.

The proportions of the different bases dissolved when igneous rocks are decomposed by weathering may be estimated roughly from the average composition of waters of the type that drain from areas of igneous rock (75 stream waters lowest in salinity and highest in silica; Clarke, 1924, p. 111, 74–79, 81, 83, 89–91, 95–97, 103–105, 107–108). After correction for "cyclic salts" (Conway, 1942, p. 135–139, 155), this is found to be, in terms of the four principal bases,

Ca	52%
Mg	11%
Na	27%
K	10%
	100%

Other bases, acid radicals, and colloidal aggregates, including significant amounts of silica and iron, in such waters may be ignored for purposes of the present estimate.

An independent estimate of the relative abundance of the principal bases in such stream water may be made by applying Conway's average percentage losses on weathering of igneous rock (Ca —67%, Mg —18%, Na —41%, and K —26%; Conway, 1942, p. 153) to Clarke and Washington's average composition of igneous rocks (Clarke, 1924, p. 29). This gives, as the expected composition of stream water from igneous areas (in per cent),

Ca	53
Mg	8
Na	25
K	14
	100

The subsequent calculations depend primarily upon the proportion of dissolved Ca to total bases, and these two estimates therefore accord closely enough for present purposes.

As decomposition of rocks proceeded, more and more dissolved bases would be carried to the sea until the solubility product of $CaCO_3$ would eventually be reached, and carbonates would begin to precipitate. Using the foregoing data, it may be found that at this stage the concentration of $[Ca^{++}]$ in the sea water would be about 0.15 mols/l, that of $[CO_3^{=}]$ about 8×10^{-6} mols/l, and the pH 5.1. Approximately 29% of the total CO_2 would then be dissolved in the sea water, 0.6% of the N_2, and 52% of the H_2S. The gases remaining in the atmosphere would have partial pressures of $P_{CO_2} = 12.2$, $P_{N_2} = 1.27$, and $P_{H_2S} = 0.28$. Summing up the quantities of dissolved acids, bases, and atmospheric gases in the sea water would give a "salinity" of about 46 g/kg. (*See* Table 2, column 2.)[4]

With further rock weathering and transportation of dissolved bases to the sea, carbonates would continue to be deposited and CO_2 subtracted from the atmosphere-ocean system. At the arbitrarily assumed $P_{CO_2} = 1.00$, at which it is conceivable that living organisms could survive, it may be found that 980×10^{20} g $CaCO_3$ would have been precipitated, leaving 53% of the original CO_2 still in the atmosphere and ocean. The concentration of $[Ca^{++}]$ would then have fallen to about 95×10^{-6} mols/l and that of $[CO_3^{=}]$ would have risen to about 13×10^{-6} mols/l. The pH would be 7.3, and the concentration of Na in the sea water would be about 12.5 g/kg. Approximately 87% of the CO_2 then

[4] The full computations would require several pages for adequate presentation; only the results are given here.

remaining in the atmosphere-ocean system would be dissolved in the sea water, 0.5% of the total N_2, and 85% of the H_2S. The gases in the atmosphere would have partial pressures of $P_{CO_2} = 1.00$, $P_{N_2} = 1.07$, and $P_{H_2S} = 0.073$. The solids and gases dissolved in the sea water would give a total "salinity" of about 67 g/kg (Table 2, column 3).

If we apply Conway's percentage losses of bases when igneous rock is weathered and use igneous rock of average composition, it may be found that approximately $17{,}000 \times 10^{20}$ g of igneous rock must have been weathered to account for the Ca, Mg, Na, and K then dissolved in sea water and previously precipitated as carbonates.

It is of interest to compare the results of these calculations with independent estimates of the quantities of some materials that have been weathered, dissolved, or precipitated during the entire course of geologic history. By methods mentioned briefly in an earlier part of this paper, the writer has estimated the total amount of carbonate rocks of all ages now remaining uneroded at 1500×10^{20} g and the total amount of igneous rock that has been weathered at $11{,}000 \times 10^{20}$ g (Table 2, columns 3 and 6).

These figures do not accord at all well with those required by the hypothesis of a dense primitive atmosphere. Even after precipitation of 980×10^{20} g $CaCO_3$, 53% of the original CO_2 would still remain in the atmosphere and ocean. To reduce this to its present small amount would mean the weathering of even more than the $17{,}000 \times 10^{20}$ g of igneous rock required at this stage and the deposition of additional C as $CaCO_3$ and organic carbon. It is thus seen that this assumption of a dense primitive atmosphere requires the weathering of more igneous rock and the deposition of more carbonates than seems likely to have been weathered and deposited in all of geologic time.

The same conclusion may be reached more di-

TABLE 2.—COMPOSITION OF ATMOSPHERE AND SEA WATER UNDER ALTERNATIVE HYPOTHESES OF ORIGIN, COMPARED WITH PRESENT-DAY CONDITIONS

	All "excess" volatiles in primitive atmosphere and ocean (Original P_{CO_2} very high)			Only a fraction of total volatiles in primitive atmosphere and ocean. (Original $P_{CO_2} \leq 1.0$ Life begins early)		Present-day conditions
	Initial stage; before rock weathering	Intermediate stage; $CaCO_3$ begins to precipitate.	Late stage; life begins at $P_{CO_2} = 1.0$	Initial stage; before rock weathering	Intermediate stage; $CaCO_3$ begins to precipitate.	
	(1)	(2)	(3)	(4)	(5)	(6)
Atmosphere (kg/cm²)	14.2	13.8	2.1	<1.1	<1.1	1.0
N_2 (% by volume)	9	9	50	7	7	78
CO_2 (% by volume)	89	89	47	90	90	0.03
H_2S (% by volume)	2	2	3	3	3	—
O_2, others (% by volume)	—	—	—	—	tr	22
Ocean ($\times 10^{20}$ g)	16,600	16,600	16,600	<990	<990	14,250
Cl, F, Br (g/kg)	18.3	18.3	18.3	18.3	18.3	19.4
ΣS, ΣB, others (g/kg)	0.8	0.8	1.3	0.1	0.1	2.8
ΣCO_2 (g/kg)	14.3	15.8	25.2	<1.1	<1.7	0.1
Ca (g/kg)	—	5.9	tr	—	<5.5	0.4
Mg (g/kg)	—	1.3	5.2	—	<1.2	1.3
Na (g/kg)	—	3.1	12.5	—	<2.9	10.8
K (g/kg)	—	1.2	4.7	—	<1.1	0.4
H (g/kg)	0.5	tr	—	0.5	tr	—
"Salinity" ‰	33.9	46.4	67.2	<20.0	<30.8	35.2
pH	0.3	5.1	7.3	0.3	5.7	8.2
$CaCO_3$ pptd. ($\times 10^{20}$ g)	None	None	980	None	None	1500± 11,000±
Igneous rock eroded ($\times 10^{20}$ g)	None	4200±	17,000±	None	<240	

rectly, without using the estimated quantities of igneous rocks eroded and carbonates deposited. The amount of Na now dissolved in sea water is known within relatively narrow limits of uncertainty. Adding any reasonable estimates that one prefers for the Na in ancient salt deposits and the Na dissolved in the interstitial water of sediments, it still comes out that this hypothesis of a dense primitive atmosphere requires the solution of much more Na in sea water, even before life began, than appears, from other evidence, to have been dissolved in all of geologic time.

Results not significantly different from these are obtained if we assume that the first rocks that were decomposed by the supposed highly acid sea water had a composition quite different from that of average igneous rock. Applying Conway's percentage losses to the weathering of average plateau basalt (Daly, 1933, p. 17) gives, as the probable relative abundance of the principal bases in stream water,

Ca	73%
Mg	12%
Na	13%
K	2%
	100%

Plateau basalt makes up only a small fraction of all the various rocks now exposed and undergoing erosion on the continents of the earth. Yet it just happens that present-day streams (Clarke, 1924, p. 119), after correction for "cyclic salts", are carrying these four bases to the ocean in proportions (per cent) similar to those that would be expected in streams from outcrops of plateau basalt:

Ca	73
Mg	11
Na	9
K	7
	100

If stream water of this composition, instead of that from areas of average igneous rock, is used in the above calculations, it may be found that, at the stage when $P_{CO_2} = 1.0$, not quite so much igneous rock needs to have been weathered ($13{,}000 \times 10^{20}$ instead of $17{,}000 \times 10^{20}$ g) but that even more $CaCO_3$ must have been deposited (1500×10^{20} instead of 980×10^{20} g). With these quantities, as with those calculated from the weathering of average igneous rock, the conclusion remains that this hypothesis of a dense original atmosphere seems to require entirely improbable amounts of rock weathering and of carbonate deposition in the early stages of earth history.

Moderate Primitive Atmosphere and Gradual Accumulation of Ocean

Let us contrast this with an alternative hypothesis that seems much more probable. The difficulties just mentioned—with what might be called the "quick-soak" hypothesis—could be avoided if, instead of all the "excess" volatiles, only a small fraction was ever present at any one time in the primitive atmosphere, and if the partial pressure of carbon dioxide never exceeded, let us say, one atmosphere. Following the same reasoning as before about the accumulation of bases in sea water, one finds that—with the alternative "slow-soak" hypothesis—carbonates would begin to precipitate when the atmosphere and ocean contained less than 10 per cent of the total "excess" volatiles. At that stage, the atmosphere would have a total pressure only one-tenth greater than at present; the salinity of the sea water would be nearly the same as that of today; and its pH would be about 5.7. The fourth and fifth columns of Table 2 show some of the conditions for two stages of this alternative hypothesis. With gradual addition of more volatiles, and as free oxygen accumulated after the advent of plant life (Van Hise, 1904, p. 956), conditions would approach closer and closer to those of today (Table 2, last column). Something of this sort seems a much more likely picture of the earth's early atmosphere and ocean.

The preceding calculation started with all the "excess" volatiles in the primitive atmosphere and ocean and followed through the consequences to be expected if the high initial pressure of CO_2 were reduced by deposition of carbonate sediments to some point where primitive forms of life could survive. This alternative hypothesis starts with the assumption that at no time did the "excess" volatiles in the atmosphere and ocean exceed such an amount that the partial pressure of CO_2 was greater than 1.0. This would permit, as seems to be required by geologic and chemical evidence (Rankama, 1948, p. 390–392, 409–414), the existence of primitive forms of life as early as, or earlier than, the deposition of the first carbonate sediments. From this

alternative hypothesis (assuming that not enough free oxygen had yet accumulated to oxidize the H_2S to SO_4) and from the previous values for the composition of incoming stream water, it is possible to estimate the conditions that would have prevailed when carbonates first began to precipitate from the sea. It may thus be found that, if P_{CO_2} never exceeded 1.0, then not more than 6% of the total "excess" volatiles could have been present in the early atmosphere and ocean; and the volume of the primitive ocean would therefore have been much smaller than at present.

At this stage the concentration of $[Ca^{++}]$ would be about 0.14 mols/1, that of $[CO_3^{=}]$ about 8.8×10^{-6} mols/1, and the pH 5.7. Approximately 3.2% of the CO_2 in the system at that time would be dissolved in the small ocean, 0.04% of the N_2, and 6.7% of the H_2S. The gases in the atmosphere would have partial pressures of $P_{CO_2} = 1.0$, $P_{N_2} = 0.077$, and $P_{H_2S} = 0.033$. The solids and gases dissolved in the sea water at this stage would give a "salinity" of about 31 g/kg (Table 2, column 5).

Poole (1941, p. 346–347, 359) has suggested that an initial supply of free oxygen, sufficient to support life and permit photosynthesis by green plants, may have been produced by photochemical dissociation of water vapor in the upper ionized layers of the atmosphere and by subsequent escape of hydrogen. In this connection, it is of interest to note that purple sulfur bacteria and certain other present-day forms can synthesize organic matter from CO_2 in an anaerobic environment that contains some H_2S. It is possible that primitive organisms of this type may have helped prepare the way for the advent of green plants on the earth (Rabinowitch, 1945, p. 4, 82–83, 99–106, 124–125). If, as, here assumed, the partial pressure of CO_2 was once as high as 1.0 atmosphere, even the rain water would have been so highly carbonated that it would leach much Fe and Si from exposed rocks, and the streams would transport significant amounts of these elements in solution (Gruner, 1922, p. 433–436; Moore and Maynard, 1929, p. 276, 293–298, 522–527; Cooper, 1937, p. 307), much as seems to have happened in parts of Precambrian time (Leith, 1934, p. 161–164).

The conditions calculated for stream water from average igneous rocks would remain essentially unchanged if we used stream water from plateau basalt instead. If, however, the partial pressure of CO_2 were less than 1.0, the fraction of total "excess" volatiles in the atmosphere-ocean system would be proportionately less than 6%; and several of the other quantities (Table 2, column 5) would likewise be decreased.

I am sure I need not warn you that the results of these calculations are not to be taken too seriously. They are based on inadequate data and laden with many "ifs and ands". Nevertheless I believe they point rather clearly to the right answer. If so, not more than a small part of the total "excess" volatiles could be residual from a primitive atmosphere and ocean. The only alternative I can think of is that these volatiles have risen to the surface from the earth's interior.

TABLE 3.—INVENTORY OF TOTAL C (AS CO_2) IN ATMOSPHERE, HYDROSPHERE, BIOSPHERE, AND SEDIMENTARY ROCKS

	$\times 10^{18}$ g	
Atmosphere	2.33	147
Ocean and fresh water	130	
Living organisms and undecayed organic matter	14.5	
Sedimentary rocks (including interstitial water)		92,000
Carbonates	67,000	
Organic C	25,000	
Coal, oil, etc.	27	

SIGNIFICANCE OF CARBON DIOXIDE IN THE ATMOSPHERE-OCEAN SYSTEM

Inventory of Carbon Dioxide

Moreover, these volatiles must have risen to the surface gradually and not in a few great bursts. Table 3 is an attempt to bring together an inventory or summary of separate estimates and thus arrive at a figure for the total carbon dioxide and carbon on and near the surface of the earth today. The present atmosphere and ocean contain approximately $1\frac{1}{2} \times 10^{20}$ grams of carbon dioxide—a small part of it in the atmosphere, somewhat more in organic matter, and most of it dissolved in water. It is significant to note that, altogether, this is less than 1 part in 600 of the total carbon dioxide and organic carbon that has, at one time or another, been in circulation in the atmosphere and ocean and is now buried in ancient sedimentary rocks.

The amount of CO_2 in the atmosphere is taken directly from Humphreys (1940, p. 81); that in sea water has been calculated as 129×10^{18} g from data on the solubility and dissociation constants in sea water of 35‰ salinity, at an average temperature of 8°C, and in equilibrium with a partial pressure of 0.0003 atmospheres of CO_2 (Harvey, 1945, p. 59, 62, 64–66). The total quantity of fresh water is here estimated at 335×10^{20} g and, from Clarke's data on average composition of stream and lake waters (1924, p. 119, 138), the amount of dissolved CO_2 is calculated to be 0.9×10^{18} g.

The quantity of CO_2 equivalent to the C in living organisms is estimated at 0.036×10^{18} g in the oceans, using data from Krogh (1934b, p. 433, 436); and at 0.029×10^{18} g on the lands, using data from Krogh (1934a, p. 421, 422), Riley (1944, p. 133), and Rabinowitch (1945, p. 6), and assuming a rough proportionality between the amount of living organic matter and the rate of carbon fixation. Undecayed particulate organic debris is estimated as equivalent to 11.8×10^{18} g CO_2 in the oceans (Krogh, 1934a, p. 422; 1934b, p. 435–436) and, very roughly, as equivalent to 2.6×10^{18} g CO_2 on the land areas, using data from Rabinowitch (1945, p. 6) on the areas of soil types and from Twenhofel (1926, p. 17) on the humus content of soils.

The estimates of carbonates and organic carbon in ancient sediments are those derived from the equations of geochemical balance mentioned above; space does not permit a more complete statement of them here. The estimates for coal, oil, etc., have been taken directly or calculated from the following sources and are here summarized in units of 10^{18} g C: Coal—7.2 (Assoc. for Planning and Regional Reconstruction, Broadsheet No. 10, 1942, p. 4); oil —0.1 (Weeks, 1950, p. 1952); oil shale—0.1; tar sands—0.025; natural gas—0.025 (Hubbert, 1950, p. 174–175); graphite—1.0×10^{-5} (U.S. reserves—Currier and others, 1947, p. 249—multiplied by 10). These total to an amount equivalent to 27.3×10^{18} g CO_2.

Carbon Dioxide Equilibria in Sea Water

If only a small part of this total buried carbon dioxide were suddenly added to today's atmosphere and ocean, it would have profound effects on the chemistry of sea water and on the organisms in the sea. These effects become more evident when we consider the important part played by carbon dioxide in the chemistry of sea water.

Table 4 is presented as a reminder of the composition of sea water and to illustrate several essential points. Sea water varies widely in its total salinity, but scarcely at all in the proportions of its dissolved constituents. The most significant exceptions are the quantities of bicarbonate and carbonate ions, which depend only in part upon salinity but are affected much more by hydrogen-ion concentration and by the amount of carbon dioxide in the atmosphere with which the water is in contact.

Note, in the column to the right, that the positive ions would exceed the negative ions by 2.38 milli-equivalents were it not that the difference between them, the so-called "excess base", is balanced by dissociated ions of carbonic and boric acid. The process of balancing the "excess base" is one of the results of the buffer action of sea water. Note also that, if the amount of sodium in sea water has been increasing continuously through geologic time, as commonly assumed, then some of the acid radicals, such as chlorine, must also have been increasing by the same amounts. Otherwise sea water would have been acid rather than alkaline throughout much the greater part of the past.

The exceedingly delicate balance that prevails between the positive and negative ions in sea water may be appreciated more fully if one considers the consequences of a decrease of only 1 part in 100 of the dissolved Na. For the salinity and temperature shown in Table 4, this would mean reducing the dissolved Na from 480.80 to 475.99 milli-equivs/l and the total of dissolved bases from 621.76 to 616.95. If the acid radicals, other than the carbonic and boric acid ions, remain unchanged, then they will exceed the bases; and the "excess base" becomes a negative quantity of $619.38 - 616.95 = 2.43$ milli-equivs/l. It may readily be shown, using the appropriate constants (Harvey, 1945, p. 59, 61–68), that under these conditions the total concentration of carbonic and boric acid ions, $C_{HCO_3^-} + 2C_{CO_3^{--}} + C_{H_2BO_3^-}$, would be only 5.6×10^{-6} milli-equivs/l, an entirely negligible quantity by comparison. Chemical balance would therefore be maintained by a concentration of $2.43 \times 10^{-3} = 10^{-2.61}$ mols or gram-ions of hydrogen per liter. That is to say, the pH of the sea water would fall from its present average value of 8.17 to 2.61.

The carbon-dioxide equilibria of sea water are numerous and complex, but related to one

another in such a way that, for a given temperature, salinity, and partial pressure of carbon dioxide, all the other variables are fixed. Published tables (Harvey, 1945, p. 59, 62, 64–66) make it possible to work out, for any given salinity and temperature, the amounts of dissolved carbon dioxide, carbonic acid, bicarbonate and carbonate ions in equilibrium with different amounts of carbon dioxide in the atmosphere, and also to find the resulting hydrogen-ion concentration in the sea water. In the left-hand diagram of Figure 1, the amounts of carbonate range from high values, which would cause precipitation of calcium carbonate, at low pressures of CO_2, to low concentrations of carbonate, which would cause solution of calcium carbonate, at high pressures of CO_2. The vertical line through the middle of both graphs represents the amount of CO_2 in the present atmosphere.

These values at low and high pressures of CO_2 would not represent stable conditions in nature. If the solubility product of $CaCO_3$ was greatly exceeded, carbonate would be precipitated from the sea water, either directly or by organic agencies, and it would continue to precipitate until equilibrium was re-established. If, on the other hand, much of the water was significantly unsaturated with $CaCO_3$, carbonate sediments lying on the sea floor would be dissolved, or Ca would simply accumulate from incoming stream water until the point of saturation was reached. From Smith's values of the solubility product for 20° and 30° C and 36‰ salinity (Smith, 1940, p. 182) and from Wattenberg's temperature and salinity coefficients (Wattenberg, 1936, p. 176), the solubility product for 8° C and 35‰ salinity may be estimated at 2.1×10^{-6}.

The right-hand diagram of Figure 1 shows, for comparison, the conditions when sea water is saturated with calcium carbonate. Note that the total carbon dioxide dissolved in sea water increases much more with increase in the partial pressure than on the other diagram.

TABLE 4.—COMPOSITION OF NORMAL SEA WATER
(Salinity 35‰, 8°C, $\rho = 1.025$, pH 8.17)

	Dissolved matter (g/kg)	Milli-mols / liter	Milli-equivalents / liter	
Cl^-	19.360	560.70	560.70	
$SO_4^=$	2.701	28.88	57.76	
Br^-	0.066	0.85	0.85	
F^-	0.001	0.07	0.07	
CO_2, H_2CO_3	0.001	0.01	—	
HCO_3^-	0.116	1.90	1.90	"Carbonate alkalinity"
$CO_3^=$	0.012	0.20	0.40	"Carbonate alkalinity"
H_3BO_3	0.022	0.35	—	
$H_2BO_4^-$	0.005	0.08	0.08	
			619.38	2.38
Na^+	10.770	480.80	480.80	
Mg^{++}	1.298	54.78	109.56	
Ca^{++}	0.408	10.46	20.92	
K^+	0.387	10.18	10.18	
Sr^{++}	0.014	0.15	0.30	
	35.161		621.76 − 619.38 = 2.38 = "Excess base"	

Recalculated for Cl = 19.37‰, $\rho = 1.025$, and pH = 8.17 from Sverdrup, Johnson, and Fleming (1946, p. 173).

Also the proportion of bicarbonate is much greater, and the pH varies much less.

These quantities of dissolved carbon dioxide in sea water and of partial pressure in the atmosphere may, if we wish, be converted into

exposed to the atmosphere in a period of from 2000 to 5000 years (Callendar, 1938, p. 224). The solid line in Figure 2 gives the pH values that would follow, perhaps more slowly, as calcium carbonate was precipitated or dis-

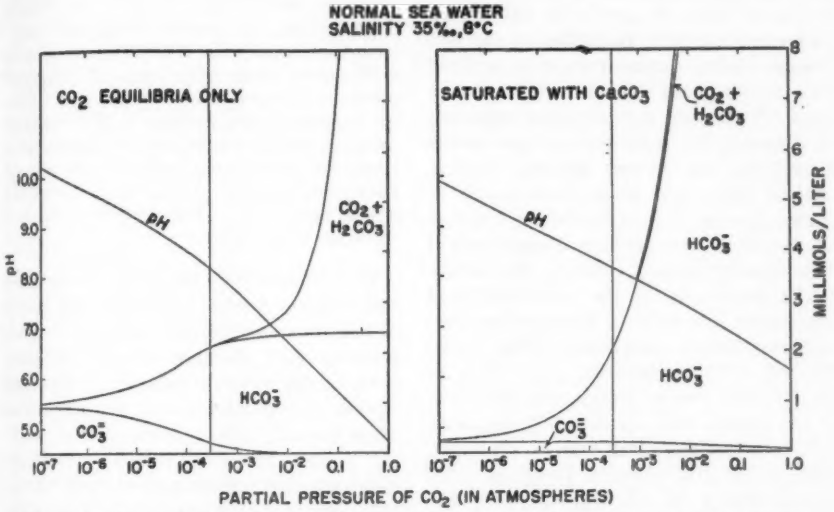

FIGURE 1.—CARBON DIOXIDE COMPONENTS AND HYDROGEN-ION CONCENTRATION OF SEA WATER AT DIFFERENT PARTIAL PRESSURES OF CARBON DIOXIDE

total carbon dioxide of the atmosphere and ocean combined. This can be done simply by multiplying through by the quantities of sea water and of atmosphere on the earth today and then adding these together. We then may see more readily the effects of changes in the total amount of carbon dioxide in the atmosphere-ocean system.

Figure 2 shows the effects that changes in the total CO_2 in the atmosphere-ocean system would have upon the hydrogen-ion concentration. Note that the dashed line gives the pH values based merely on the carbon-dioxide equilibria and that these are the values that would follow relatively soon, geologically speaking, after sudden changes of total carbon dioxide. The rate at which sea water could re-establish equilibrium after a change in atmospheric CO_2 depends mainly on the rate at which the deeper water is brought into contact with the air. From what is known about oceanic circulation, it has been estimated that the entire volume of sea water is solved. Figure 3 shows the effects that changes in the total CO_2 would have upon the partial pressure of CO_2 in the atmosphere. The values of total CO_2 shown in Figures 2 and 3 are the original ones before the adjustment that would be caused by solution or precipitation of $CaCO_3$.

The relationships shown in Figures 2 and 3 are for the present-day volume of sea water, a salinity of 35‰, and a temperature of 8° C. The effects of reducing the volume of sea water, of decreasing the salinity, or of raising the temperature are to shift the lines for pH in Figure 2 and for P_{CO_2} in Figure 3 somewhat to the left; but the essential relationships are unchanged.

Effects on Organisms

What would be the effects of such changes in total carbon dioxide on organisms living in the sea? We have already considered the narrow tolerance ranges of some animals for total salinity and for composition of the waters

in which they live. For many forms of life, the concentration of bicarbonate and hydrogen ions and the carbon-dioxide tension are among the most critical factors in their chemical environment. A number of higher marine animals (the herring, for example) are extremely sensitive to small changes in the pH of their environment. A large proportion of the eggs of some marine animals remains unfertilized if the acidity of sea water departs more than about 0.5 pH from normal. Lower organisms are commonly less sensitive; but many species of mollusks, sea urchins, Medusa, diatoms, bacteria, algae, and others seem unable to tolerate a range of more than about 1 unit of pH. Recent workers attribute a larger part of these observed biologic effects to the carbon-dioxide tension or to the concentration of bicarbonate ions than to hydrogen-ion concentration directly (McClendon, 1916, p. 148; Shelford, 1918, p. 101–102; Gail, 1919, p. 288, 295, 297; Powers, 1920, p. 381–382; 1939, p. 73; Atkins, 1922, p. 734–735; Legendre, 1925, p. 213; Singh Pruthi, 1927, p. 743; Valley, 1928, p. 215–216, 218–220; Davidson, 1933; Rogers, 1938, p. 97, 285, 286, 294, 430, 653–656, 679, 680; Edmondson, 1944, p. 43–45, 63, 64; Allee and others, 1949, p. 175, 197).

Having thus reviewed in outline the part played by carbon dioxide in the chemistry of sea water and in the environment of marine organisms, let us now return to the inventory of carbon dioxide and to the significance of the fact that more than 600 times as much of it is buried in ancient sedimentary rocks as there is now in circulation in all the atmosphere, hydrosphere, and biosphere.

If only one-one hundredth of all this buried carbon dioxide were suddenly added to today's atmosphere and ocean (that is, if the amount in the present atmosphere and ocean were suddenly increased $\left(\frac{600}{100} + 1 = \right)$ sevenfold, from 1.3×10^{20} to 9.1×10^{20} g), it would have profound effects on the chemistry of sea water and on the organisms living in the sea. The first effect would be to change the average pH of sea water from about 8.2 to 5.9 (Fig. 2, *a*). This acid water would be much less than saturated with calcium carbonate, and thus further changes would follow. Eventually, when equilibrium was re-established, the partial pressure of carbon dioxide in the atmosphere would be about 110 times its present value, and the pH of the sea water would end up at an average of about 7.0 (Fig. 2, *b*).

The two values of pH, 5.9 and 7.0, are those for present volume and salinity of sea water and a mean water temperature of 8° C. For three-fourths of the present volume of sea water but with present salinity and temperature, these values are 5.8 and 6.9, respectively. For a salinity of 27‰ instead of 35‰ but with present volume and temperature of water, the corresponding values are 5.8 and 7.1. For a mean water temperature of 30° C but with present volume and salinity of sea water, these values are 5.9 and 6.8.

The effects of these changes on living organisms would be drastic. If the supposed increase of carbon dioxide happened suddenly, it would probably mean wholesale extinction of many of the marine species of today. If, however, the increase were gradual, so that organisms could adapt themselves by generations of evolutionary changes, the effects would be much less disastrous—but perhaps no less clearly recorded in the physiological adaptations of the surviving forms. From the paleontologic record it appears improbable that any change so drastic as an abrupt sevenfold increase of carbon dioxide has happened, at least since the beginning of the Cambrian.

Effects on Composition of Sediments

The conclusion that the amount of carbon dioxide in the atmosphere and ocean cannot have varied widely through much of the geologic past does not rest solely on the narrow bicarbonate and pH tolerances of many organisms. The mineralogical and chemical compositions of sedimentary rocks tell much the same story.

Carbon dioxide is constantly being added to the present atmosphere and ocean by the weathering of limestones and other rocks, by artificial combustion (Callendar, 1940, p. 399), and from volcanoes, geysers, and hot springs. At the same time, it is also being subtracted by the deposition and burial of calcium carbonate and of organic carbon in sediments.

Let us, for the moment, ignore the contribu-

tions of carbon dioxide from artificial combustion and from volcanoes and hot springs, and compare only the rates of addition by rock weathering and of subtraction by sedimentation. These rates are not known accurately, of course, but from several lines of evidence it appears that the loss by sedimentation must exceed the gain from weathering by something like 10^{14} grams of carbon dioxide each year. At this rate of net loss, the total carbon dioxide in the atmosphere and ocean would be reduced to about one-fourth its present value in one million years, and the concentration of hydroxyl ions in sea water would be so high that magnesium hydroxide—the mineral brucite—would be precipitated and almost completely take the place of calcite as a common marine sediment. Even if liberal allowance is made for

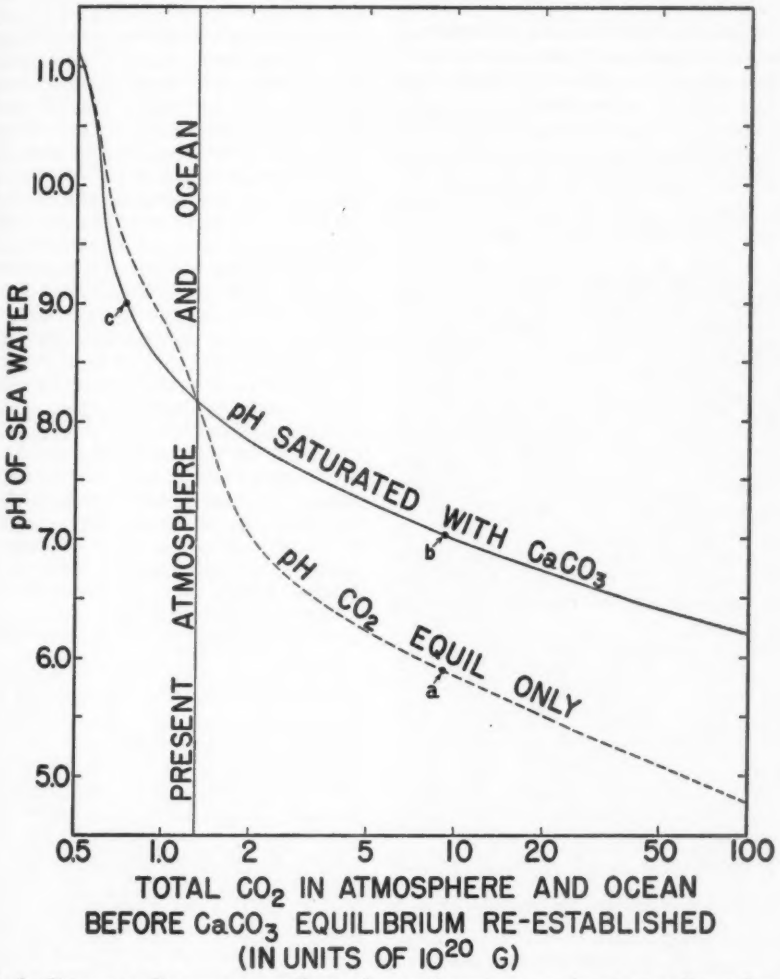

FIGURE 2.—EFFECTS OF CHANGES IN THE TOTAL CARBON DIOXIDE IN THE ATMOSPHERE-OCEAN SYSTEM UPON THE HYDROGEN-ION CONCENTRATION OF SEA WATER (AT 35‰ SALINITY AND 8° C)

a probable slowing down of weathering and of the synthesis of organic matter as carbon dioxide is subtracted from the atmosphere-ocean system, it would still be only about 2 million years until brucite would largely take the place of calcite deposition. No such occurrences of sedimentary brucite are known to me in rocks of any age, and it appears unlikely that this condition was ever reached in the geologic past. It seems much more likely that the net losses by sedimentation have been roughly balanced by a source neglected in the foregoing calculation—that is, by continuous additions of carbon dioxide from volcanoes and hot springs.

It seems clear that the loss of CO_2 from the atmosphere-ocean system by sedimentation must, in the long run, somewhat exceed the gain from rock weathering, but the amount of the difference is uncertain. Sedimentary rocks contain, on the average, roughly equivalent amounts of CaO and CO_2 (Clarke, 1907a, p. 169; 1907b, p. 269; 1924, p. 30). Their decomposition by weathering or precipitation by sedimentation does not disturb this equivalent ratio significantly. But igneous rocks contain many times more CaO than CO_2 (Clarke, 1924, p. 29). Thus the proportion of igneous rocks to all rocks undergoing erosion at any time is a rough measure of the amount of CO_2 that must be supplied from some other source in order to maintain the observed ratio in newly formed sediments. Several different methods of estimation (based on the probable quantities of CO_2, organic C, and Ca in rocks undergoing erosion, in material being transported by streams, and in sediments being deposited, both today and in the geologic past) yield values for the net loss of CO_2 from the atmosphere-ocean system of from 0.3×10^{14} to 4.0×10^{14} g/yr. The estimates of 1×10^{14} to 2×10^{14} g appear somewhat the more reliable, and, for purposes of the following calculations, a value of 1×10^{14} g CO_2/yr has been adopted.

At this rate of net loss, the total CO_2 would be reduced from 131×10^{18} to 31×10^{18} g in $\frac{(131-31) \times 10^{18}\ \text{g}}{1 \times 10^{14}\ \text{g/yr}} = 1{,}000{,}000$ years, and the pH of sea water would then be about 9.0. Wattenberg and Timmermann (1938, p. 87–88) found the solubility product of $Mg(OH)_2$ to be 5×10^{-11} in sea water; and Harvey remarks that "when the hydrogen-ion concentration of sea water falls and the pH rises above circa pH 9, magnesium hydroxide separates as a precipitate with calcium carbonate" (1945, p. 25). If CO_2 subtraction continues after the precipitation of $Mg(OH)_2$ has begun, the relative amounts of Mg and Ca in sea water and the various CO_2 equilibria cause $Mg(OH)_2$ to accumulate in far greater amounts than $CaCO_3$.

This estimate of the time required to reduce the total CO_2 in the atmosphere and ocean to the point when brucite would be deposited neglects several factors—one that would operate to decrease the time and two that would operate to increase it. The several chemical relations involved require that, as the pH of sea water is raised from 8.2 to 9.0, 70 mg/l of $CaCO_3$ must, on the average, be subtracted by precipitation to maintain the various equilibria. With the present volume of sea water, this would mean the deposition as carbonate of about 43×10^{18} g of CO_2 in addition to the average net subtraction of 10^{14} g/yr based on present conditions. The total subtraction of (131−31=) 100 units of CO_2 is thus made up in part of 57 units lost because of normal sedimentation. If this is not replaced by CO_2 from other sources, this primary loss causes the precipitation, as carbonate, of 43 additional units of CO_2 in order to maintain the various equilibria. Figure 2 shows the effects of primary rather than secondary changes of total CO_2 in the system; and point *c* (pH 9.0) is therefore plotted at (131−57=) 74 units instead of at 31 units. For purposes of the present calculation, we are interested in the time required for the primary loss of 57 units of CO_2. That is, if there were no other corrections to be considered, the time required to reduce the pH to 9.0 would be only

$$\frac{(131 - 31 - 43) \times 10^{18}\ \text{g}}{1 \times 10^{14}\ \text{g/yr}} = 570{,}000 \text{ years.}$$

The total rate of loss of CO_2 by sedimentation is the rate of deposition of carbonates and that of organic carbon. Over the years, the mean rate of carbonate deposition depends upon the rate at which streams bring new Ca to the sea, and this in turn depends upon the rate of rock weathering on the lands. It seems reasonable to assume that the rate of chemical weathering is controlled in large measure by the partial pressure of CO_2 in the atmosphere (Van Hise, 1904, p. 465, 476; Clarke, 1924, p. 110–111) and by the quantity of decomposing organic matter (Jensen, 1917, p. 255–258, 267–268) which likewise depends upon atmospheric CO_2, as is discussed somewhat more fully below. As the atmospheric CO_2 would decrease under the assumed conditions of this calculation, the rate of rock weathering and hence the rate of carbonate deposition would likewise decrease.

Similarly the rate of accumulation and burial of organic carbon is probably roughly proportional to the rate at which new organic matter is produced, both on the lands and in the sea. This in turn depends largely if not entirely on the rate of carbon

fixation by photosynthesis which, at optimum light intensities, varies almost directly with the partial pressure of CO_2, up to a P_{CO_2} of about 3, 4, or 5 times that of the present (Brown and Escombe, 1905, p. 40–41; Hoover, Johnston, and Brackett, 1933, p. 10–17; Rabinowitch, 1945, p. 330–331). Thus as the partial pressure of CO_2 decreased under the assumed conditions, the rates of deposition of both carbonate sediments and organic carbon would probably decrease in roughly the same proportion.

From the relationship between CO_2 and P_{CO_2}, it may be found that, when the total CO_2 is reduced to 31×10^{18} g (Fig. 3, *d*, plotted at 74×10^{18} g to show the effects of primary loss), the P_{CO_2} would be reduced to only one-forty-fourth its present value. Numerical integration indicates that, over this range, the P_{CO_2} would average about 28% of its present value. Thus the mean rate of CO_2 deposition would be only about 0.28×10^{14} g/yr. Combining the effects of these several corrections, a revised estimate of the time required to reduce the total CO_2 to the point where $Mg(OH)_2$ would begin to precipitate would be $\frac{(131 - 31 - 43) \times 10^{18} \text{ g}}{0.28 \times 10^{14} \text{ g/yr}} =$ 2,000,000 years. This is a relatively brief interval in geologic time, and it seems necessary to conclude that CO_2 must have been supplied to the atmosphere-ocean system more or less continuously from some extraneous source in order to account for the absence of brucite as a common marine sediment. The foregoing calculations are based on the constants for sea water of normal salinity at a mean temperature of 8°C. The calculated values are somewhat different if other salinities and temperatures are assumed but not enough so to modify this conclusion significantly.

Possible variation in the amount of CO_2 in the atmospheres of the past might be estimated, if sufficiently reliable data were available on the rate at which organic carbon has been deposited during different geologic periods. As stated above, the amount of organic carbon that becomes buried along with sediments and thus removed from circulation at any time is probably roughly proportional to the amount of it then in existence as organic matter. This in turn depends, through a narrow but significant range, on the partial pressure of atmospheric CO_2. The rate of deposition of organic carbon is thus a rough measure of the partial pressure of CO_2 in the atmosphere. For this reason, it is tempting to suppose that the CO_2 content of the atmosphere may have been much greater than at present during those parts of the Carboniferous, Cretaceous, and Tertiary periods when great quantities of coal and other organic deposits accumulated.

Available data do not support this possibility, however. To begin with, the total amount of carbon in all coal, petroleum, oil shale, etc., is negligible in comparison with the carbon that is present (to the extent of only a few tenths of 1 per cent) in the far more abundant ordinary sedimentary rocks (Table 3). Furthermore, local deposits of highly organic sediments do not necessarily mean unusually large accumulations at that same time over the entire earth. Finally, even widespread deposits of highly organic sediments are not in themselves sufficient evidence; the time elapsed during accumulation of the sediments is an essential factor in determining whether or not the rate of carbon deposition was abnormally rapid.

Merely to list the data needed for this calculation is sufficient to demonstrate how far we now are from being able to appraise at all accurately, from this type of evidence, the rates of burial of organic carbon through the geologic past. One needs (a) reliable information on the organic carbon content of different types of sediment of different ages, (b) reasonably good estimates of the relative amounts of different sediment types deposited during a given interval of time, and (c) data on the number of years required for deposition. Of these several categories of essential data, the third is probably now known better than the other two.

Nevertheless, something of interest may still be obtained, even from such data as now exist. The average composition of the principal types of sedimentary rocks (Clarke, 1924, p. 30, 547, 552) affords one of the bases for estimation; the relative abundance of sandstone, shale, and limestone in rocks of different ages may be taken from such compilations as Schuchert's (1931, p. 49); and the estimated durations, in millions of years, of the different periods are available from several sources (for example, Holmes, 1947b, p. 144). When these various estimates are combined, along with estimates of the amounts of contemporaneous deep-sea sediments, one finds no indication whatever that the rate of carbon burial has ever exceeded that of today. In fact this evidence, taken at face value, appears to show that the rate of carbon accumulation has been increasing gradually and rather uniformly ever since the beginning of geologic time.

This apparent evidence of an increasing rate of carbon deposition needs to be viewed with considerable skepticism, however. It may readily be recognized as simply one other aspect of the widely observed relationship that the maximum thicknesses of sedimentary rocks deposited during a unit of time appear to increase progressively as one ascends through the geologic column. This general relationship has been noted and variously interpreted by a number of geologists. Gilluly (1949, p. 574–582)

has clearly pointed out that it does not necessarily mean any real increase in the rate of sedimentation with time but may be explained equally well as the result of the accidents of preservation and exposure of ancient sediments. Whatever may be the correct explanation of this commonly observed relationship, one finds no indication from this type of evidence that the average rate of deposition of organic carbon was greater at any time in the past than it is today.

Possible Limits of Variation in the Past

Even if the rate of carbon deposition, averaged over fairly long intervals of time, has remained approximately constant through much of the past, it still seems likely that the rate would have varied somewhat from this average, as the result of irregularities of volcanic eruptions and other possible sources of supply. The observed fact that many plants are capable of a higher rate of carbon-dioxide assimilation and photosynthesis, if the carbon-dioxide content of the atmosphere is several times greater than normal, is perhaps most simply explained as the result of the adaptation of ancestral forms to an atmosphere that was at times somewhat richer in CO_2 than at present. If we adopt this interpretation and take a P_{CO_2} of about 0.0015 (or five times that of the present) as an optimum condition for photosynthesis (Brown and Escombe, 1905, p. 40–41; Hoover, Johnston, and Brackett, 1933, p. 15–16), we have a basis for estimating what may have been the total amount of CO_2 in the atmosphere-ocean system at such times.

With the present volume and salinity of sea water, a mean water temperature of about 8° C, and a prevailing P_{CO_2} of 0.00030, the quantity of CO_2 in the atmosphere and ocean is calculated to be 2.33×10^{18} and 129×10^{18} g, respectively. If, with the other conditions remaining constant, the total quantity of CO_2 in the atmosphere and ocean were increased from 131×10^{18} to 157×10^{18} g, the various CO_2 equilibria would distribute the CO_2 between atmosphere and water so that the P_{CO_2} would rise to the assigned value of 0.00150 (Fig. 3, *e*) and the pH would fall to 7.5. This sea water would then be much less than saturated with $CaCO_3$, and, as a result, $CaCO_3$ would, on the average, be dissolved from the ocean floor until equilibrium was re-established. At that stage, the total quantity of CO_2 in the atmosphere and ocean would be about 177×10^{18} g; the pH would rise to 8.0 (nearly its original value), and the P_{CO_2} would fall back to about 0.00059 (Fig. 3, *f*).

It would probably give a more reasonable estimate of the limiting conditions to assume that the arbitrary P_{CO_2} of 0.00150 prevailed *after* rather than before adjustmsnt to $CaCO_3$ equilibrium, which would mean that the P_{CO_2} was at times well above that limit. On this alternative assumption, the total quantity of CO_2 would have increased from 131×10^{18} to 212×10^{18} g, the addition having come from some unspecified source. The CO_2 equilibria alone would distribute this total between the atmosphere (49×10^{18} g) and sea water (163×10^{18} g) so that the P_{CO_2} would rise to 0.00630 (Fig. 3, *g*) and the pH would fall to 6.9. Sea water of this acidity would dissolve $CaCO_3$ from the ocean floor until about 62×10^{18} g of CO_2 had been added to the system, thereby bringing the total to $(212 + 62 =) 274 \times 10^{18}$ g. At that stage, when equilibrium was re-established, the quantity of CO_2 in the atmosphere and ocean would be about 11.6×10^{18} and 262×10^{18} g, respectively; the P_{CO_2} would fall back to the assigned value of 0.00150 (Fig. 3, *h*); and the pH would end up at 7.8. It is worth nothing that, although the quantity of CO_2 in the atmosphere would have increased by a factor of $\frac{11.6}{2.33} = 5$, the quantity in the ocean would have increased by only $\frac{262}{129} = 2$.

The volume and salinity of sea water and the temperature of the water have been held constant in the foregoing calculation. The probable effects of their variation may be estimated by similar calculations. If the volume of the ocean were only three-fourths as great as at present, the quantity of CO_2 dissolved in sea water would be only 194×10^{18} instead of 262×10^{18} g when the P_{CO_2} was 0.00150. For salinity ranges of as much as 20 per cent above and below those of present average sea water, the net effects on the distribution of CO_2 between atmosphere and sea water are almost negligible. If the mean water temperature were 30°C instead of 8°C, the quantity of CO_2 dissolved in sea water would be about 138×10^{18} instead of 262×10^{18} g when the P_{CO_2} was 0.00150. Thus the effects of each of these other controlling variables is such

that, if the P_{CO_2} were 5 times that of today, the total CO_2 in the atmosphere-ocean system would probably not have been much more than about twice the present total of 131×10^{18} g.

photosynthesis, which in turn is proportional to the P_{CO_2} through this range. This would mean that, as the P_{CO_2} increased by a factor of 5, the quantity of CO_2 equivalent to organic carbon in the biosphere

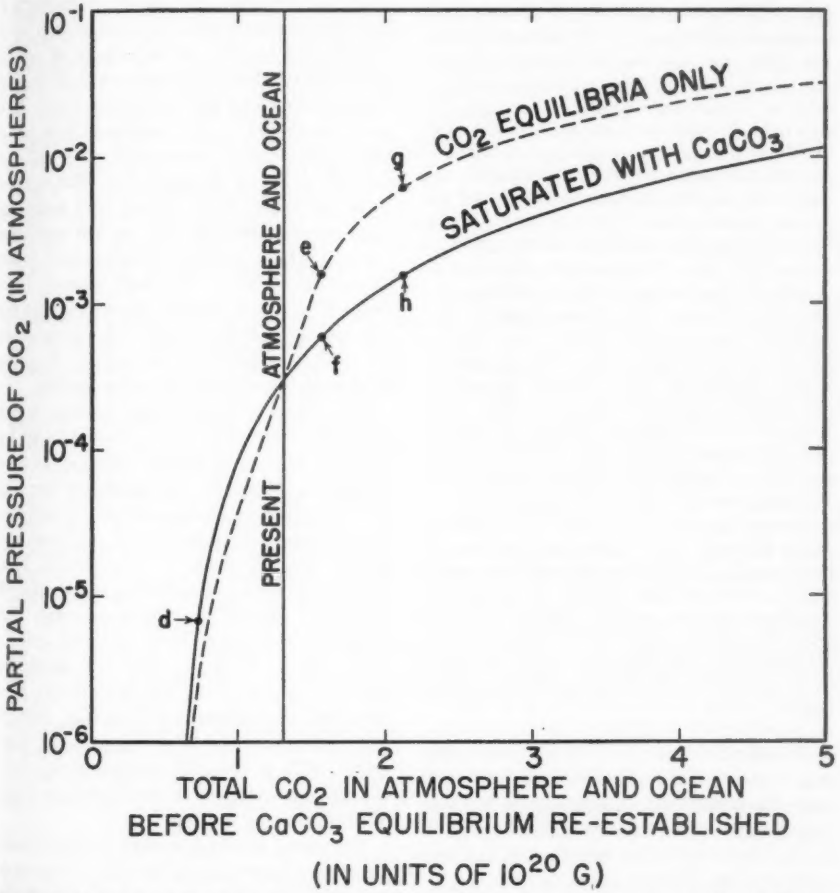

FIGURE 3.—EFFECTS OF CHANGES IN THE TOTAL CARBON DIOXIDE IN THE ATMOSPHERE-OCEAN SYSTEM UPON THE PARTIAL PRESSURE OF CARBON DIOXIDE IN THE ATMOSPHERE (AT 35‰ SALINITY AND 8° C)

All these calculations have considered only the amounts of CO_2 in the atmosphere and ocean and have ignored any changes in the total quantity of living organisms and undecayed organic matter in the biosphere, as the P_{CO_2} increased from 0.00030 to 0.00150. By an extension of the reasoning followed above, in the section entitled Effects on Composition of Sediments, it might be estimated that the quantity of organic matter in the biosphere at any time would be roughly proportional to the rate of

would also have increased from the estimated 14.5×10^{18} g of today (Table 3) to about 72.5×10^{18} g. This would bring the total quantity of C in circulation in the atmosphere, hydrosphere, and biosphere under these limiting conditions to about $(274 + 72 =) 346 \times 10^{18}$ g CO_2. A similar "correction" might be made for organic C in the biosphere in each of the other calculations. None of the conclusions would be modified, however, and the basis

for the "correction" itself seems too uncertain to warrant this refinement.

The possible effect of variation in the CO_2 content of the atmosphere upon world climate has been discussed by meteorologists and others for many years. The hypothesis, developed in some detail by Arrhenius (1896) and T. C. Chamberlin (1897, p. 680; 1899), was that an increase of atmospheric CO_2 would act as a trigger mechanism, retarding radiation of heat from the earth, and thereby starting a train of events that would cause general warming. Physical measurements subsequently indicated that the heat absorption by water vapor far outweighs that by CO_2, and the hypothesis fell into general disfavor (reviews in Clarke, 1924, p. 52–53, 147–149; Humphreys, 1940, p. 584–586, 621–622). Recent work, however, seems to have reopened the question to some extent (Callendar, 1938, p. 231–232; 1941; Dobson, 1942; Elsasser, 1942, p. 64). Callendar's later estimates (1938, p. 231; 1941, p. 32) call for an increase of 20 to 30 fold over the present atmospheric CO_2 in order to account for the warmest climates indicated by Tertiary floras, and they call for a decrease to about one-tenth the present content (corresponding to a decrease to about 0.4 of the present total CO_2 in atmosphere and ocean) to account for the coldest epochs of continental glaciation. Other lines of evidence, discussed elsewhere in this paper, make it seem unlikely that the P_{CO_2} has been 20 to 30 times greater than at present, at least in later parts of geologic time. But it appears possible, if CO_2 has the climatic effects some have claimed for it, that decreases in the amount of atmospheric CO_2 may have been a contributory cause of some of the epochs of glaciation.

From several lines of evidence, it seems difficult to escape the conclusion that, for a large part of geologic time, carbon dioxide has been supplied to the atmosphere and ocean gradually and at about the same rate that it has been subtracted by sedimentation.

Sources of Supply

If the bulk of the carbon dioxide in the atmosphere, hydrosphere, biosphere, and sedimentary rocks cannot have been residual from a primitive atmosphere, one must look for other sources of this gradual and continuous supply. At first glance, the possibility seems tempting that this unidentified source may have been the great reserve of carbonates and organic carbon in the ancient sediments. Crustal movements bring these older sedimentary rocks to the surface, where they undergo weathering and gradually release their contained carbon to the streams and atmosphere. If sedimentary deposits accumulated only on the continents and continental shelves and if no igneous rocks were eroded, this problem of the supply of carbon would not concern us, for the quantities of older sedimentary rocks eroded from and new sediments deposited on the continents have probably maintained a reasonably close balance through much of geologic time. But igneous rocks have also been eroded, and sediments have accumulated not only on the continents but also in large quantity in the deep-sea basins (Twenhofel, 1929, p. 394, 400; Kuenen, 1937; 1941; 1946). It is evident, therefore, that the weathering of sedimentary rocks alone could not suffice to maintain the supply of carbon dioxide to the atmosphere-ocean system.

If weathering of sedimentary rocks is not sufficient, what about the carbon released by metamorphism or by actual melting of the older sediments? Many of the successive steps in metamorphism of siliceous carbonate rocks involve progressive decarbonation of the mineral species (Bowen, 1940, p. 245, 256–257, 266); and actual melting of sediments would drive off all carbon dioxide or carbon monoxide that could not be dissolved in the resulting melt. In many areas, volcanic rocks and associated schists of Precambrian age have largely been converted into carbonates (Van Hise, 1904, p. 972; Collins and Quirke, 1926, p. 31–32; Macgregor, 1927, p. 159–162), possibly as the result of high concentrations of carbon dioxide driven off from near-by sources.

The extent to which ancient sediments have undergone decarbonation by metamorphism or melting is not known. It seems very improbable, however, that such processes are adequate to account for the continuous supply of carbon dioxide to the atmosphere and ocean. As an outside estimate, we may assume that *all* Precambrian and *half* of all Paleozoic sedimentary rocks now remaining uneroded on the continents have undergone complete (not merely partial) decarbonation and that *all* the resulting carbon dioxide escaped into the atmosphere-ocean system. This would reduce the estimated amount of CO_2 now buried in sedimentary

rocks from 920 × 10^{20} (Table 1) to about 780 × 10^{20} g and would mean that the amount now buried is about 530 (instead of 600) times as great as that in circulation in the present atmosphere, hydrosphere, and biosphere. In other words, even if decarbonation of sedimentary rocks has been far more extensive than seems likely, the essential problem of accounting for this huge discrepancy still remains.

It may be worth noting also that, judged by the estimated relative abundances shown in Table 1, the volatile constituents that could be released on complete decarbonation, dehydration, etc., of average sedimentary rock would be very different in over-all composition from the "excess" volatiles.

One other possible source of carbon dioxide may be worth mentioning. Stony and iron meteorites contain an average of 0.04 per cent and 0.11 per cent carbon, respectively (Brown, 1949b, p. 626). Conceivably the supply of carbon dioxide to the earth's surface might have been replenished continuously by showers of meteoritic dust from interplanetary or interstellar space (Van Hise, 1904, p. 970, 973, 974). A brief consideration, however, of the exceedingly slow rate at which meteoritic debris is reaching the earth today (less than 1 mm/billion years—Jeffreys, 1933—to about 10 cm/billion years—Nininger, 1940, p. 461) and the apparent absence of geologic evidence that the rate has been significantly higher any time since sedimentary rocks began to accumulate makes it seem highly improbable that extra-terrestrial sources have made important contributions to the earth's supply of carbon dioxide since very early in the history of the planet.

The conclusion, arrived at so laboriously here, that carbon dioxide has been supplied to the atmosphere and ocean gradually and at about the same rate that it has been subtracted by sedimentation, is not particularly new. The Swedish geologist, A. G. Högbom, clearly recognized in 1894 (summarized in Arrhenius, 1896, p. 269–273) that carbon dioxide has been supplied continuously to the earth's surface since early in geologic history and that the chief source of this supply must have been volcanic or juvenile. Essentially this same conclusion has been restated many times (T. C. Chamberlin, 1897, p. 654–656; Fairchild, 1904, p. 97, 109–110; Clarke, 1924, p. 58–59; Macgregor, 1927, p. 156–157; Jeffreys, 1929, p. 312; Goldschmidt, 1934, p. 415; 1938, p. 101; Eskola, 1939; Conway, 1943, p. 170–174; Cotton, 1944; Hutchinson, 1944, p. 180, 192; 1947, p. 300–301); but it is a significant fact about the history of the earth that has sometimes been ignored and, as such, it deserves frequent repetition.

Similarity of "Excess" Volatiles and Magmatic Gases

The evidence bearing on the probable source and rate of supply of carbon dioxide has been emphasized in the preceding discussion because the chemical and physiological effects of carbon dioxide are relatively well known. Perhaps for this reason, carbon dioxide appears to be of particular significance in our understanding of the history of the atmosphere and ocean. Comparable data on the effects of other volatiles are apparently not available; but, from scattered information, approximately the same conclusions are suggested for the source and rate of supply of chlorine, oxygen, and several other constituents of the atmosphere and hydrosphere.

In summarizing the evidence on the source of carbon dioxide, Hutchinson put it succinctly when he stated that "it seems unreasonable to accept juvenile addition in the case of one constituent, and, in the absence of very strong reasons, to deny it in the case of the others" (1944, p. 180). A number of other writers have reached much the same conclusion (Fairchild 1904, p. 98, 103; Evans, 1919; Jeffreys, 1929, p. 147, 312; Eskola, 1932b, p. 66–67, 69–70; Lane, 1932, p. 318; Gilluly, 1937, p. 440–441; Goldschmidt, 1938, p. 17, 24, 101; Wildt, 1939, p. 143; R. T. Chamberlin, 1949, p. 253, 255–256). (*See also* references regarding the earth's loss of inert gases in section on Dense Primitive Atmosphere-Solution in the Melt.)

Let us now return to the "excess" volatiles we started with and see how they compare in composition with gases that are escaping from volcanoes, fumaroles, and hot springs and with gases that are occluded in igneous rocks. Considering that these gases have come in contact and presumably reacted with diverse rock

TABLE 5.—VOLUME PERCENTAGES OF GASES FROM VOLCANOES, ROCKS, AND HOT SPRINGS

	Volcano gases from Kilauea and Mauna Loa* (26 samples)			Gases from rocks: Basaltic lava and diabase* (13 samples)			Gases from rocks: Obsidian, andesitic lava, and granite* (17 samples)			Gases from fumaroles of the Katmai region† and from steam wells and geysers of California and Wyoming** (23 samples)			
	Minimum	Maximum	Median	Minimum	Maximum	Median	Minimum	Maximum	Median		Minimum	Maximum	Median
CO_2	0.87	47.68	11.8	0.89	15.30	8.1	0.08	20.26	2.0	CO_2	0.03	1.24	0.02
CO	0.00	3.92	0.5	0.02	8.28	0.2	0.01	2.22	0.5	CO	—	0.01	tr
H_2	0.00	4.22	0.4	0.38	6.18	1.2	0.08	11.60	0.4	O_2	0.00	0.08	tr
N_2	0.68	37.84	4.7	0.27	7.21	2.0	0.03	3.90	1.2	CH_4	0.00	0.30	0.11
A	0.00	0.66	0.2	0.00	0.04	tr	0.00	0.02	tr	H_2	0.00	0.29	0.15
SO_2	0.00	29.83	6.4	—	—	—	—	—	—	$N_2 + A$	0.00	0.31	0.02
S_2	0.00	8.61	0.2	0.08	1.96	1.1	0.00	2.89	0.2	NH_3	—	0.02	0.01
SO_3	0.00	8.12	2.3	—	—	—	—	—	—	H_2S	0.00	0.10	0.02
Cl_2	0.00	4.08	0.05	0.06	1.33	0.5	0.01	10.59	0.5	HCl	0.01	0.57	0.06
F_2	—	—	—	0.00	14.12	3.8	0.25	7.80	2.3	HF	0.00	0.10	0.03
H_2O	17.97	97.09	73.5	71.32	92.40	83.1	69.44	98.55	92.9	H_2O	98.04	99.99	99.58
			100.0			100.0			100.0				100.00

* Analyses by E. S. Shepherd (1938, p. 321, 326).
† Allen and Zies (1923, p. 126, 142).
** Allen and Day (1927, p. 76; 1935, p. 87).

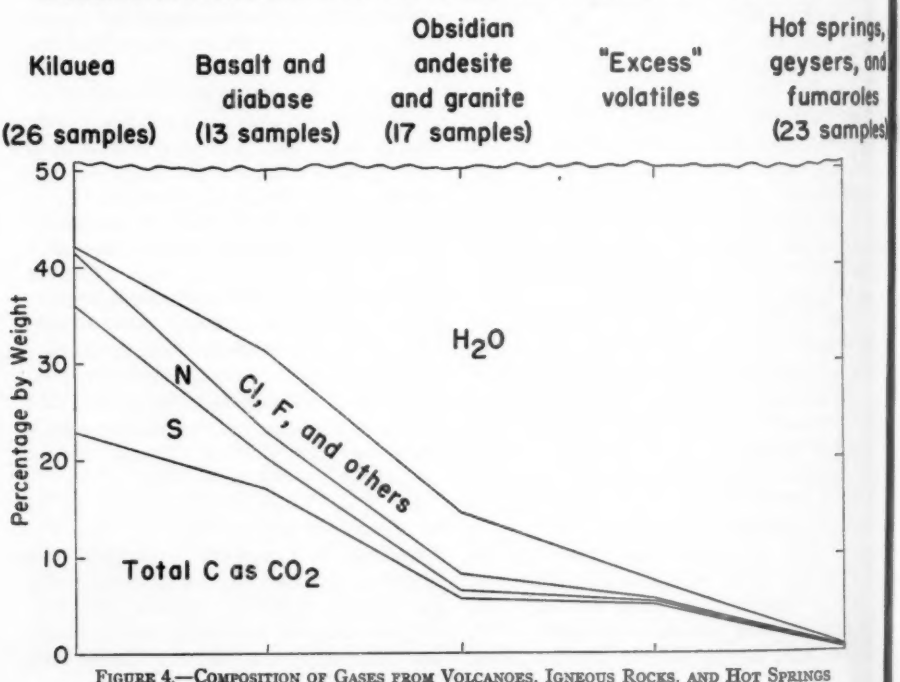

FIGURE 4.—COMPOSITION OF GASES FROM VOLCANOES, IGNEOUS ROCKS, AND HOT SPRINGS COMPARED WITH "EXCESS" VOLATILES

types and ground waters and that they probably contain at least some recirculated or *resurgent* volatiles that have been absorbed from earlier sedimentary rocks (Shepherd, 1938, p. 312-313), it is to be expected that they will be

TABLE 6.—COMPOSITION OF GASES FROM VOLCANOES, IGNEOUS ROCKS, AND HOT SPRINGS, AND OF "EXCESS" VOLATILES IN ATMOSPHERE, HYDROSPHERE, AND SEDIMENTARY ROCKS

(Median analyses, recalculated from volume to weight percentages)

	Kilauea and Mauna Loa	Basalt and diabase	Obsidian, andesite, and granite	Fumaroles, steam wells, and geysers	"Excess" volatiles
H_2O	57.8	69.1	85.6	99.4	92.8
Total C as CO_2	23.5	16.8	5.7	0.33	5.1
S_2	12.6	3.3	0.7	0.03	0.13
N_2	5.7	2.6	1.7	0.05	0.24
A	0.3	tr	tr	tr	tr
Cl_2	0.1	1.5	1.9	0.12	1.7
F_2	—	6.6	4.4	0.03	tr
H_2	0.04	0.1	0.04	0.05	0.07
Others	—	—	—	—	tr
	100.04	100.0	100.04	100.01	100.04

variable in composition. The extreme ranges of composition of these gases are in fact wide; but median compositions by groups indicate consistent differences in the gases in rocks and lavas of different types (Tables 5 and 6). If we compare these median gas compositions with the relative abundance of the constituents of the "excess" volatiles, we see that an average of the gases in granitic rocks and those from hot springs and fumaroles approximates closely the over-all composition of the "excess" volatiles (Fig. 4). Judged solely by compositions then, it is conceivable that all the hydrosphere and atmosphere may have come from such gases (Allen, 1922, p. 39–42, 52). Several questions must be answered, however, before we may consider such a conclusion.

ESCAPE OF VOLATILES DURING CRYSTALLIZATION OF MAGMAS

Is there some reasonable mechanism by which these gases from the interior could be the source of the "excess" volatiles? In the cooling and gradual crystallization of complex silicate melts, certain minerals crystallize out first and thus enrich the remaining melt in other constituents, such as silica, alumina, and the alkalies (Bowen, 1928, p. 293, 297–298) and water, carbon dioxide, and other volatiles (Allen, 1922, p. 57). As crystallization proceeds, the concentration of the remaining volatiles rises and therefore also their vapor pressures. As Morey (1922; 1924) and others (Goranson, 1931, p. 499; Ingerson, 1940, p. 784) have pointed out, when these vapor pressures become greater than the weight of overlying rocks, escape to the surface may be violent, as in volcanic eruptions; or where the crystallizing melt lies at greater depths, the volatiles may escape more gradually along fractures in the overlying rocks.

How adequate is such a mechanism to explain the quantity of volatiles that must have risen to the earth's surface in the course of geologic time? Probably the total quantity of extrusive rocks that has risen to the earth's surface in all of geologic time, although large, is still insufficient to furnish more than a few per cent of the water of the ocean, unless extrusive rocks bulk larger than the total of all sedimentary rocks and unless they originally contained more than 10 per cent of water. (*See also* Verhoogen, 1946, p. 746.) Intrusive rocks, on the other hand, appear to be a more adequate source. Goranson's experimental work has shown that a melt of granitic composition can dissolve as much as 9.3 per cent of water at 900° C and at a pressure equivalent to a depth of 15 kilometers (Goranson, 1931, p. 493). The amount of water in natural magmas, however, is a subject about which petrologists are not agreed. Gilluly's estimates, which are fairly representative of others[5], are that the water content of many magmas has been "fully 4 per cent for basalts and perhaps as much as 8 per cent for deep granites" (1937, p. 441).

Let us estimate 4 per cent water in an average magma. Then, as igneous rocks now contain an average of about 1 per cent water, this would mean that magmas have commonly given off about 3 per cent of water during crystallization.

[5] In a paper that appeared after this was written, Ingerson (1950, p. 813–814) has reviewed the estimates made by a number of geologists. Gilluly's estimates lie near the middle of the entire range.

On these assumptions, the crystallization of a shell of igneous rock 40 kilometers thick would account for the entire volume of water in the ocean. In a similar calculation, Gilluly (1937, p. 440–441) estimated that the ocean could have been derived from the escape of 5.8 per cent of water from the sialic rocks of the crust.

Hot Springs as Possible Channels of Escape

How might this quantity of volatiles have escaped to the earth's surface? Hot springs are much less spectacular than volcanoes, but they are more numerous and more widely distributed, and they may be the principal channels by which volatiles escaping from crystallizing magmas have reached the surface. For example, Allen and Day (1935, p. 40, 87; Allen, 1935, p. 6, 12; Day, 1939, p. 328–329) have estimated, from the temperature of escaping steam in Yellowstone Park, that the hot springs of that area are discharging ground water that has been heated by mixing with about 10 or 15 per cent of magmatic steam[6].

Hot springs are much more abundant in the area of the Idaho batholith than in that of the much younger lava flows of the adjacent Snake River Plains (Meinzer, 1924, p. 296–297; Stearns, Stearns, and Waring, 1937, Fig. 11, p. 82). More than 100 hot springs and groups of hot springs are known within the limits of the Idaho batholith, and the discharge and temperature of a number of them have been measured (Stearns, Stearns, and Waring, 1937, p. 136–151). If these springs are heated by the same process of mixing of ground water and magmatic steam that is believed to operate in the Yellowstone, then one may estimate roughly that the average percentage of magmatic water is about 2½ per cent. The Idaho batholith is thought to be of Cretaceous age (Ross, 1936, p. 382–383), yet it is still discharging hot waters. Can it be that batholiths cool much more slowly than is generally supposed? If so, this may be one explanation for the long-continued and relatively uniform supply of volatiles that seems to be indicated by the geologic record.

The average temperature of the hot springs of Idaho that have been measured is about 39°C, and the mean annual temperature there is near 8°C. Making the same assumptions about original temperature and mixing as those used in the similar calculation that follows, we find that 1 g of magmatic steam would be sufficient to heat 35.5 g of ground water to 39°C. This would mean 2.7 per cent magmatic water in the final hot-spring mixture.

A crude estimate of the total quantity of water being discharged today from all the hot springs of the earth may be made from available data on the springs of a smaller area. Stearns, Stearns, and Waring (1937, p. 115–191) list more than 1000 thermal springs or groups of springs within the limits of the continental United States. Undoubtedly some of these are heated by deep artesian circulation, but probably only a small proportion, for many are located in areas of igneous rock. To allow roughly for such artesian heating, the hot springs of the eastern two-thirds of the United States are here omitted from calculation. If the total area of the States, as thus "corrected", is a fair sample of the continents and if the oceans contribute about half as much per unit area, then the annual discharge of all hot springs, multiplied by 3 billion years, gives a total volume somewhat more than 100 times that of the present ocean. That is to say, if hot springs are delivering to the surface an average of only 0.8 per cent of juvenile water, they could, in the course of geologic time, account for the entire volume of the ocean. This is not counting any contribution of water directly from volcanic eruptions.

Stearns, Stearns, and Waring consider only springs with water more than about 8°C above local mean annual temperature. Of the 1059 hot springs they list for the entire United States, 721 show estimated discharges, and 616 show water temperatures. The total discharge of the 721 measured or estimated springs is 511,600 gal/min. The 338 springs for which no discharge is recorded include the 96 hot springs and geysers of Yellowstone

[6] Rubey and Murata (1941) have investigated a small but apparently typical group of hot springs farther south along the Wyoming-Idaho border. From relations between the compositions of water and of contained gases and the temperatures at different outlets, it appears that approximately 5 per cent of the water of these springs is of magmatic origin. For estimates of the amount of magmatic water in other areas, *see* Zies (1929, p. 73–74) and White and Brannock (1950, p. 573).

National Park, for which an estimate of 49,400 gal/min is available from another source (Allen, 1935, p. 5). Estimating separately by States for the unmeasured springs, it appears that a total of approximately 800,000 gal/min is a fair approximation for the discharge of all 1059 springs. Comparing the water temperatures with estimated mean annual temperatures and weighting the differences in proportion to the quantities discharged, one arrives at the estimate that the springs are heated, on the average, about 20°C above mean annual temperature.

By the same methods, the hot springs of the western one-third of the country (Ariz., Calif., Colo., Idaho, Mont., Nev., N. Mex., Ore., Utah, Wash., and Wyo.) yield a total of approximately 700,000 gal/min or 1.4×10^{18} g/yr of water. Taking the area of all continental U. S. as one-thirtieth that of all the continents and using the discharge from only these 11 western States would give a total from all continents of 42×10^{18} g H_2O/yr. The rate of hot-spring discharge on the ocean floor is unknown. Presumably it is lower than that on the continents but, to judge by the evidence of extensive lava flows there (Pettersson, 1949, p. 186), it is probably considerable. If the mean rate per unit area from the oceans is taken as half that from the continents, then the hot-spring discharge from the continents and oceans combined is about 66×10^{18} g H_2O/yr. This rate, continued over a period of 3 billion years, the assumed age of the earth (Holmes, 1947a; Bullard and Stanley, 1949; Ahrens, 1949, p. 254, 257–258; H. N. Russell, 1949, p. 11, 12, 15, 20, 24), would yield a total of 2.0×10^{28} g H_2O. The $16{,}600 \times 10^{20}$ g of "excess" water (Table 1) would be 0.0083 of this amount.

It is of some interest to compare this rough estimate of 0.8 per cent juvenile water with another estimate arrived at in quite a different way. Judged by those measured, the hot springs in the 11 western States have an average temperature of about 29° C, which is roughly 20° above the local mean annual temperatures. For purposes of calculation, let us assume that these springs are heated by deep-seated magmatic steam at an original mean temperature of, let us say, 600° C, and that this steam is cooled to 29° by mixing with ground water at the mean annual temperature of 9° C. In cooling from 600° to 29°, steam and water would release about 1100 gram-calories for each gram of magmatic water (estimated from data in Keenan and Keyes, 1936, p. 79–81). This would be sufficient to heat $\left(\frac{1100}{20} = \right)$ 55 grams of ground water from 9° to 29°. On these assumptions, the hot springs of the United States are being heated by mixing with an average of $\left(\frac{1}{56} = \right)$ 1.8 per cent of magmatic water.

It would, of course, be rash to attach any great significance to the numerical results of these calculations, for the difference between the two estimates is subject to several possible explanations. Nevertheless, it may be worth reminding ourselves that the estimate of 0.8 per cent is for new *juvenile* water and the 1.8 per cent estimate is for total *magmatic* water of whatever source. It is conceivable that the difference of 1.0 per cent between the two estimates may be a rough measure of the *resurgent* water, not newly risen from the earth's interior for the first time but recirculated water that magmas have resorbed from earlier sedimentary rocks—"second-hand" water that is now being returned to the surface.

Generation of Local Magmas

Even if we should grant all that has gone before about the probable juvenile origin of the "excess" volatiles, there still remains the question of why the magmas that released the volatiles have not long since cooled off and crystallized completely. In spite of seismic evidence that indicates the possibility, at least in some areas, of a layer of molten rock at a depth of about 80 km (Gutenberg and Richter, 1939b; Gutenberg, 1945, p. 302–307), serious difficulties restrain us from assuming a world-encircling layer (Daly, 1946, p. 712, 721–722) of slowly crystallizing magma from which volatiles have been escaping continuously ever since the world began. However, the abundant field evidence of local intrusive masses, some of them very large, that have been generated repeatedly throughout the past suggests a possible answer.

If we are willing to adopt the eclectic method of borrowing from various theories those parts of each that we like best, then a possible mechanism suggests itself—a mechanism by which local masses of hydrous magma might be generated more or less continuously throughout geologic time. The geothermal gradient varies

from place to place (Van Orstrand, 1939, p. 132–141; Landsberg, 1946); but probably below a depth of 50 to 100 kilometers rocks are heated to significantly higher temperatures (Holmes, 1915, p. 111; Adams, 1924, p. 468–472; Jeffreys, 1929, p. 154; Gutenberg, 1939, p. 162; Urry, 1949, p. 176; or one may calculate his own thermal gradient from revised data on the radioactive generation of heat in different rock types—Keevil, 1943, p. 299; Birch, 1950a, p. 612–613, 616, 619), temperatures at which they would melt, if the confining pressures were not so great (Holmes, 1916, p. 269; Adams, 1924, p. 462, 468–472; Anderson, 1938, p. 50–51, 56–57, 72–73; Benfield, 1940, p. 157–158; Buddington, 1943, p. 139).

Under these conditions, several processes may operate to cause local melting. As sediments accumulate in a region, their low thermal conductivity makes them act as a blanket to impede the upward flow of heat (Jeffreys, 1930, p. 328). And in geosynclinal troughs, where sediments accumulate to great thicknesses, the entire column of underlying rocks is depressed into zones where the temperature is significantly higher (Eskola, 1932a, p. 456–457, 468–469). Temperatures may thus be reached at which some of the minerals in the deeper rocks are melted. Moreover, as strains accumulate in the earth's crust and the rocks eventually fail by fracture, sudden localized relief of pressure and shearing stresses may, without increasing the temperature, bring some of the minerals above their melting points (Johnston and Adams, 1913, p. 210–223, 248–253; Johnston and Niggli, 1913, p. 602–603, 613–615; Bowen, 1928, p. 314; Holmes, 1932, p. 556; Wahl, 1949, p. 148). As has been shown by Bowen (1928, p. 311–320), Eskola (1932a, p. 474), and others (Holmes, 1926, p. 315–316; 1932, p. 545, 556; Kennedy, 1938, p. 38–39; Buddington, 1943, p. 132–133, 137–139; Wahl, 1949, p. 161–163), such selective fusion of the lower-melting minerals would form magmas more granitic and at the same time richer in volatiles than the original rock.

Fusion would cause increase of volume, and the resulting hydrous, granitic magmas would tend to rise, partially recrystallizing as they moved into higher, cooler zones. This process would "pump" nearly all the volatile materials from the original rock; in recrystallizing these magmas could form suites of intermediate and granitic rocks; and the volatiles would largely escape to the earth's surface. By this mechanism, the volume of granitic rocks would increase near areas of sedimentation and where the crust was for any other reason unstable. As these unstable areas migrated with the continental margins and with the sites of mountain making, the granitic rocks would thus increase progressively in volume (Eskola, 1932a, p. 468–469, 480; Lawson, 1932, p. 358–361; Wilson, 1950, p. 101, 108, 111; Bucher, 1950, p. 500–504; Hurley, 1950).

Corollaries and Tests of the Suggested Mechanism

Isostatic Considerations

This leads to an important corollary of the suggested mechanism. If granitic rocks have grown progressively in volume through geologic time, then isostatic equilibrium would require more or less continuous readjustments, by which the thickening blocks of lighter, granitic or continental types of rocks would rise higher and higher above adjacent blocks of undifferentiated material. In another terminology, this would mean that the ocean basins must sink deeper as the continental blocks thicken. If the ocean basins have sunk at approximately the same rate as the volumes of granitic rocks and of sea water have grown, then it is no coincidence, but the effects of one single process, that the surface of the sea has oscillated back and forth near the same level on the continental shelves throughout the geologic past.

Isostatic balance between the oceanic and continental blocks affords a basis for estimating the relative quantities of water and granitic rocks and of the original subcrustal matter from which they may have been derived by selective fusion. The area of the earth's surface that rises above sea level is roughly 1.49×10^{18} cm^2; but the actual margins of the continental blocks lie much farther oceanward, beyond the continental shelves and near the base of the steep continental slopes. If 3000 meters below sea level is taken as approximately the base of the continental slopes, then the total area of the continental blocks is about 2.32×10^{18} cm^2 (Sverdrup, Johnson, and Fleming, 1946, p. 15, 19, 21).

The mean elevation of the continental blocks, as thus defined, is about 0.12 km above sea level, and the mean depth of the ocean basins (below the −3000 m contour) is about 4.6 km below sea level. The average density of sea water *in situ* is approximately 1.04 (Sverdrup, Johnson, and Fleming, 1946, p. 219). From independent estimates, mentioned briefly elsewhere in this paper, it appears that the mean thickness of deep-sea sediments may be about 1.3 km and their mean density, when filled with interstitial water, about 2.3. For purpose of this calculation, the average density of rocks to several tens of kilometers below the continents (the sedimentary, "granitic", and "intermediate" shells) may be taken as 2.9, and that of undifferentiated ultrabasic rocks to similar depths below the deep-sea sediments as 3.2.

From these several estimates, the average depth of compensation below the continents may be found to be

$$0.12 + \frac{(0.12 \times 2.9) + 4.6\,(3.2 - 1.04) + 1.3\,(3.2 - 2.3)}{3.2 - 2.9} = 38.3 \text{ km.}$$

This value accords reasonably well with the average depth to the Mohorovičić discontinuity, which is found widely distributed beneath the continents but is not recognized with certainty over extensive areas beneath the oceans (Macelwane, 1939, p. 237–239; Gutenberg and Richter, 1939a, p. 321–322). This major discontinuity is often interpreted as representing the approximate base of the crustal materials that make up the continents.

It seems permissible to assume that these densities of continental and oceanic materials have remained approximately the same throughout the past. It follows then that, if there was a time when the mean depth of the ocean basins and the mean thickness of deep-sea sediments were half their present values, isostatic equilibrium would require that the mean elevation of the continents and the mean depth of compensation below the surface of the continental blocks would, at that time, also be just half their present values. If the total areas of the continental blocks and of the oceanic basins have remained approximately the same (peripheral growth of the continents having approximately balanced subtraction by continental foundering), this strictly proportionate deepening of the oceanic basins and thickening of the continental crust, which is required by isostatic balance, would accord with the suggested mechanism of selective fusion which likewise would require that the volumes of sea water and of continental rocks have grown in strict proportionality.

The total mass of the continental blocks today may be estimated from the results of the previous calculation at $(2.32 \times 10^{18}\ \text{cm}^2)\ (38.3 \times 10^5\ \text{cm}) \times 2.9\ \text{g/cm}^3 = 258 \times 10^{25}$ g. Allowing for the probable quantity of crystalline rocks that has been eroded from the continents in the past and is now represented as deep-sea sediment on the ocean floor, one obtains an estimate of 2.65×10^{25} g for the total mass of crustal rocks that have been formed on and in the continental blocks. It is to be noted that this estimate makes no allowance for any similarly differentiated crustal rocks that may now be lying foundered beneath the sea.

According the the eclectic hypothesis here suggested, this 2.65×10^{25} g of crustal rock has been formed, largely by selective fusion and subsequent recrystallization of the resulting magma, from an unknown quantity of subcrustal rock materials. The total mass of the earth's silicate mantle that lies directly below the continents, to a depth of 2900 km and above the earth's "iron core," may be estimated at about 185×10^{25} g. Much of this subcrustal matter may still be in very nearly its original condition (Birch, 1950b). But that part of it lying above the maximum depth (of about 700 km) to which deep-focus earthquakes are recorded (Gutenberg and Richter, 1941, p. 4–11; Benioff, 1949, p. 1844–1854) may have been worked over and subjected, at one time or another during the course of geologic history, to at least some selective fusion. The total mass of this subcrustal matter below the continental areas and between depths of 38 and 700 km is about 50×10^{25} g.

On the assumption that this is roughly the quantity of subcrustal matter that may have yielded both the "granitic" magmas that became the continental rocks and the volatile substances that have formed the atmosphere and ocean, it is of interest to compare the relative amounts of each. With an average water content of 1.15 per cent (Clarke, 1924, p. 29), the 2.65×10^{25} g of present crustal rocks would contain 3050×10^{20} g H_2O. This, added to the $16{,}600 \times 10^{20}$ g of "excess" H_2O in the present atmosphere, ocean, and sedimentary rocks, gives a total of $19{,}650 \times 10^{20}$ g H_2O that has escaped, along with the 2.65×10^{25} g of "granitic" magmas, from the subcrustal matter. Thus the original $(50 + 2.65) \times 10^{25}$ g of subcrustal matter would have yielded, by selective fusion, $\frac{2.65}{52.65} = 5.0$ per cent, or 1 part in 20, of magmas that have risen to form the crustal rocks and $\frac{0.1965}{52.65} = 0.37$ per cent H_2O. If the "granitic" magmas originally held all this water in solution, they would have

contained at that time an average of $\frac{0.1965}{2.65} = 7.4$ per cent H_2O, of which $\frac{0.166}{2.65} = 6.3$ per cent subsequently escaped to the surface during recrystallization of the magmas.

These estimates are, of course, only crude ones based upon very uncertain foundations. They make no allowance for possible "granitic" rocks now foundered beneath the oceanic depths; to whatever extent any continental rocks now lie beneath the ocean floor, the calculated percentages of water in the "granitic" magmas and of water lost from the subcrustal matter must be reduced proportionately.

A rough but independent check on the reasonableness—or otherwise—of this estimate of the amount of subcrustal matter that has been selectively fused can be made by considering certain substances that may have been concentrated in the crustal rocks or at the surface as the result of the supposed differentiation. In the absence of better information, the average composition of the original undifferentiated peridotitic material may be taken as similar to that of stony meteorites (Bowen, 1928, p. 315–316; 1950. *See also* references in section on Dense Primitive Atmosphere—Solution in the Melt). The silicate phase of stony meteorites contains an average of 0.063% H [equivalent to 0.56% H_2O] and 0.199% K [equivalent to 0.24% K_2O] (Brown, 1949b, p. 626). Thus the calculation that 0.37% H_2O has escaped from the subcrustal matter to a depth of 700 km would mean that, on the average, $\frac{0.37}{0.56} = 66\%$ of this mass of material has undergone selective fusion.

If the residual unmelted fraction remaining after selective fusion has an average composition similar to that of dunite, it would contain about 0.04% K_2O (Daly, 1933, p. 20). The average igneous rocks now exposed on the continents contain 3.13% K_2O (Clarke, 1924, p. 29). From these average compositions, we may calculate that the original subcrustal matter has yielded about $\frac{(0.24 - 0.04)}{(3.13 - 0.04)} = 6.5\%$, or 1 part in 15, of magmas that have crystallized to form the present crustal rocks. This ratio of 1 part of magma to 15 parts of subcrustal matter, when compared with the previous estimate of 1 part in 20 (based on total material to 700 km), would mean that on the average about 75% of the mass of subcrustal matter to a depth of 700 km has undergone selective fusion.

It is only fair to add that ratios of much more and much less than 1 part in 15 may be obtained by using for comparison other elements than K and other rock types (peridotite instead of dunite and diorite instead of the average igneous rock now exposed at the upper surface of the continental blocks). Yet when one considers that the process of selective fusion would act unequally on different constituents of the subcrustal matter and would probably remove much larger proportions of the readily volatile substances, such as water, than of the more refractory rock-forming minerals, it is surprising to find even such rough agreements as those just mentioned. About all that may safely be concluded is that the general order of magnitude of the ratio of "granitic" magmas to subcrustal matter appears not unreasonable.

Other Geologic Tests

The least one can ask of any hypothesis is that it be consistent with the known facts. But it is not always as easy as it sounds to apply this test, because much information and hard work are sometimes required to formulate corollaries of a hypothesis so that they can be tested by the known facts. I have thought of a few geologic tests of this suggested mechanism, but I am sure that many others have not occurred to me.

At first thought, the widespread epicontinental seas of the past seem to contradict the suggestion flatly. But areas are not to be confused with volumes. If the ocean basins have been sinking relative to the continental blocks, then one must look largely to the ocean floor, rather than to the continents, for evidence of a growing volume of sea water. Unfortunately, our information about the ocean floor is still very sketchy. The flat-topped sea mounts, or guyots, discovered by Hess (1946) indicate a considerable sinking of the ocean floor, some part of which may mean a general rise of sea level and not merely local deformation. Other lines of evidence, such as submerged terraces and great thicknesses of shallow-water limestones on some oceanic islands (Carsey, 1950, p. 376), suggest, although likewise they do not prove, the same conclusion. It has long been recognized that very few areas that once were part of the deep sea have later risen to become part of the continents (Walther, 1911, p. 60–61). Furthermore, the truncation of structural trends at some continental margins and the geographic distribution of certain animals and plants seem to require that parts of some land masses have broken off and foundered to oceanic depths

(Barrell, 1927, p. 303–305; Moore, 1936, p. 1786). These are all rather obvious, general tests that yield results consistent with, but certainly not proving, an increasing volume of ocean basins through geologic time.

Undoubtedly other and more rigorous tests should be looked for in the stratigraphic and structural record on the continents, where many more facts are available than we yet know about the ocean basins. A gradually increasing volume of sea water and of marine sediments would mean a gradual rise of sea level and relatively slow marine transgressions, modified of course by the effects of local deformation, and interrupted by relatively rapid withdrawals of the sea, when the ocean floor sank to new positions of equilibrium. Such sudden withdrawals of the sea would affect parts of all continents at the same time (Moore, 1936, p. 1803–1805, 1808). On the other hand, if hydrous granitic magmas have formed by the mechanism indicated, then mountain-making movements, as contrasted with emergences of the continental shelves, need not have been even approximately contemporaneous from one continent to another (Gilluly, 1949).

How well or how poorly do the geologic facts bear out such corollaries as these? Or if these are not valid tests of the suggested mechanism, I trust that others will think up much better ones.

As I warned at the beginning, this subject ramifies almost endlessly into many problems of earth history. A satisfactory solution of the problem of the source and history of the earth's air and water depends on the solution of a great many other questions. Because it is so closely related to many others, this problem is likely long to remain one of those hardy perennials that need a new look and reappraisal every few years, as new observations accumulate. Perhaps what I have presented here is simply another case of putting 2 and 2 together and getting 22, instead of 4. But this review and these tentative conclusions will have served their purpose if they help stimulate—or if they provoke—critical observations and critical thinking about the history and significance of sea water.

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U. S. GEOLOGICAL SURVEY, WASHINGTON 25, D. C.
MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, MAY 10, 1951.
PUBLISHED WITH THE PERMISSION OF THE DIRECTOR, U. S. GEOLOGICAL SURVEY.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, P. 1149 SEPTEMBER 1951

Short Notes

OUTCROP OF PAYETTE FORMATION ALONG SNAKE RIVER NEAR HAGERMAN, IDAHO

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The Payette formation of Miocene age crops out in the Snake River Valley from Boise to the Idaho-Oregon boundary and is exposed also in the foothills bordering the south side of the Snake River Valley in southern Idaho for many miles east of Boise.

However, no outcrops of this formation have been found previously in Idaho along Snake River from the Boise area eastward due to the widespread veneer of Pleistocene basalts and Hagerman lake beds of Pliocene age (Lake Idaho deposits).

Recently the writer found an outcrop, about 40 feet square, of firmly cemented yellowish cross-bedded arkosic sandstone containing scattered pebbles up to 2 inches across of quartzite, chalcedony, and black limestone in the northwest corner of the SW¼ of sec. 29, T.8 S., R.14 E., B.M., Twin Falls County, on the west side of a secondary road branching off the paved highway from Hagerman to Buhl, Idaho. The outcrop lies near the mouth of Salmon Falls Creek about 1¼ miles northwest of the Banbury Hot Springs in an area mapped by the writer (1938, Pl. 5) as a fault complex of Hagerman lake beds and Banbury-type basalt near the Riverside volcanic cone of Pliocene age. The sandstone strikes N.20° E. and dips 14° NE. It is overlain unconformably by 75 feet of Hagerman lake beds and associated basalt.

No fossils were found to establish the outcrop positively as Payette, but its lithology clearly distinguishes it from the weakly cemented later sediments in this area.

It appears that the Pliocene and Pleistocene formations are thin at this place, probably as a result of a horst or earlier rocks being exposed by erosion.

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HOPE, IDAHO